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DIGGING FOR WORMS: CHARACTERIZING NEAR-GROUND VORTICITY
DURING SUPERCELL TORNADOGENESIS WITH DOPPLER RADAR AND
LIDAR

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DIGGING FOR WORMS: CHARACTERIZING NEAR-GROUND VORTICITY
DURING SUPERCELL TORNADOGENESIS WITH DOPPLER RADAR AND
LIDAR

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SCHOOL OF METEOROLOGY

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Dedication

My master's thesis is dedicated to my grandfather, or "Papa", Mike Frost, who passed away this February. My Papa is the best man I've known, and the biggest supporter of my pursuit of a career in atmospheric science. Papa taught me the importance of hard work and never giving up. Through my 6 years of challenging classes and late nights working on assignments and research, I always remembered Papa saying "Robby, if you do what you love, you'll never work a day in your life". I would not be the man I am today without him, I love you Papa.

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Abstract

The question of why some supercells produce tornadoes and some do not has motivated decades of research by the severe storms community. Advancing the understanding of the dynamical processes behind tornado formation is thought to be beneficial for improving tornado forecasts and warnings. The current understanding of tornadogenesis involves four stages, the first being the formation of a rotating updraft called a mesocyclone. The second is the generation of rotation near the ground, which can be converged and amplified by the vertical motion beneath the mesocyclone. A recent numerical modeling study by Markowski (2024) proposes the idea that the inflow, not the outflow, can act as the origin of near-ground vertical vorticity for supercell tornadogenesis. This inflow vertical vorticity is associated with turbulent coherent structures in the surface layer of the planetary boundary layer (colloquially referred to as “vorticity worms”), which generate organized streaks of vertical vorticity often greater than 0.05 s^{-1} in highly sheared environments. This finding is contrary to nearly half a century of tornado research, which has pointed to the tilting of horizontal vorticity by a supercell’s downdrafts as the origin of surface rotation for tornado formation. Because vorticity worms exist in the typically clear-air inflow of supercells, they are difficult to capture in most radar observations. Additionally, vorticity worms are confined to the surface layer, only spanning a vertical depth of a few hundred meters, adding additional observational challenges.

In the current study, we turn to observations of the 10 May 2010 Central Oklahoma tornado outbreak by the OU-PRIME radar, which was a polarimetric C-band system and was located in the inflow of the Moore-Choctaw EF4 tornado. The narrow beamwidth, high sensitivity, and use of a 0.25° elevation angle provide a rare opportu-

nity to characterize near-ground (< 100 m AGL) vertical vorticity in all regions of the Moore-Choctaw supercell during tornadogenesis. We use the 2D multi-level wavelet transform to estimate the wavelength of inferred vertical vorticity from the azimuthal shear of radial velocity (ζ_{inf}). Dividing the near-field of the supercell into inflow and outflow partitions, we find that wavelengths $200 < \lambda < 800$ m are the most common at ≈ 100 m AGL in both regions. The wavelet analysis also confirms that the near-field inflow is predominantly composed of vorticity worms while the outflow is made up of vortex patch-like features (e.g., lacking preferred orientations). Partitioned time-height series of circulation from a dual-Doppler analysis (DDA) with OU-PRIME and KTLX show similar values between the inflow and outflow. Additionally, we inspect vertical vorticity flux into the inflow and outflow partitions, finding positive flux into the near inflow and negative flux into the outflow partition during the entire period. We speculate that this is due to the orientation of the storm-relative winds near the mesocyclone, which are southeasterly in the inflow and exceed 50 m s^{-1} during tornadogenesis. In contrast, the storm-relative winds in the RFD are very weak and have an easterly component in some regions. This orientation could indicate the inflow winds are better oriented to advect vertical vorticity towards the mesocyclone in this case, though the 2+ minute volumetric updates of this dataset prevent us from directly tracking this process. Vertical profiles of wind speed are constructed in the inflow, revealing a shallow layer (≈ 50 m deep) of substantial streamwise horizontal vorticity. Lastly, we subjectively identify vorticity worms in the inflow and discuss their stretching potential in the context of the Fischer et al. (2024) tornadogenesis model.

Though our analysis is insufficient to answer if vorticity worms directly facilitated tornadogenesis in this case, our findings highlight the inflow region as a potentially large source of vertical vorticity for tornado formation. Discussion of other supercell cases confirms that vorticity worms are not unique to our case or the Markowski (2024)

simulations. Lastly, we discuss environmental parameters that may be important in the context of vorticity worms and present ideas to further investigate the influence of turbulent coherent structures on supercell tornadogenesis.

Chapter 1

Introduction

Decades of research has been dedicated to understanding the intricate dynamics of supercell tornado formation due to their immense danger to life and property (Anderson-Frey and Brooks, 2019). However, the severe storms community still struggles to answer one of the most studied questions in meteorology:

Why do some supercells produce tornadoes and some do not?

1.1 The Current Understanding of Supercell Tornadoogenesis

Advances in computing and radar technologies in recent years have allowed the community to better understand the fine-scale details of tornadoogenesis. A recent review article by Fischer et al. (2024) outlines the community's current understanding of supercell tornadoogenesis, breaking the process into four stages. The first stage is the acquisition of rotation by the low-level mesocyclone, which most notably occurs through the tilting of environmental horizontal vorticity created by vertical wind shear (Rotunno, 1981; Lilly, 1982; Davies-Jones, 1984). This tilting is gradual and lifts parcels away from the ground, generating little vertical vorticity near the ground (Davies-Jones and Markowski, 2013). Because of this, some other mechanism must generate vertical vorticity near the ground for a tornado to form (Fischer et al., 2024). This is the second

stage, and Fischer et al. (2024) primarily discusses the formation of vertical vorticity along downdrafts from the supercell itself. These are often referred to as “vorticity rivers” or “vortex patches”, and are typically seen along a supercell’s rear-flank gust front (RFGF), forward-flank convergence boundary (FFCB), or left-flank convergence boundary (LFCB) (Dahl et al., 2014; Orf et al., 2017; Parker, 2023).

The third stage is the organization and stretching of this near-surface vertical vorticity. Vertical vorticity formed near the ground by storm-scale processes is rarely tornado strength, so some mechanism must intensify this rotation. This is why the first stage is so important, as a rotating low-level mesocyclone creates negative pressure perturbations that enhance the strength of the updraft, which converges and stretches vorticity to tornado strength (Wicker and Wilhelmson, 1995). The success of this stage depends on a strong, persistent, and well-organized updraft below the lifting condensation level (LCL), as well as favorable cold pool characteristics (Markowski and Richardson, 2009, 2014; Coffey and Parker, 2017). The fourth stage is the actual transition to a tornado, marked by the formation of a tornado boundary-layer (TBL) and corner flow region. The intensifying vortex’s low pressure drives strong inflow air that sharply turns upward into an updraft, typically in the core of the tornado (Fischer and Dahl, 2022). At this stage, tilting of large horizontal vorticity in the corner flow is sufficient to maintain the tornado. It is important to note that this tilting is much more abrupt than that explained in stage 1, allowing for the generation of vertical vorticity at the surface.

While this understanding of tornadogenesis is convincing and is based on numerous observational and numerical studies, it strongly emphasizes the generation of vertical vorticity by storm downdrafts as the primary source of near ground rotation for tornado formation. However, vertical vorticity is present in the clear-air planetary boundary layer (PBL) when sufficient vertical wind shear and vertical motions are present in the atmospheric surface layer (ASL), which is typically the lowest 10% of the PBL.

Nonetheless, many previous numerical simulations of supercells lack fully turbulent PBLs in their environment (Klemp and Wilhelmson, 1978; Wilhelmson and Wicker, 2001; Rotunno, 2013). In addition, many studies employed a free-slip lower boundary condition until the most recent decade (Coffer and Parker, 2015; Dahl et al., 2014; Markowski and Richardson, 2014; Orf et al., 2017; Schueth et al., 2021), which does not simulate friction near the ground. Another factor that has limited the analysis of PBL turbulence in supercell inflow is the difficulty of clear-air retrievals by Doppler weather radar (Bodine et al., 2023). The environmental inflow of supercells is often rain-free, offering fewer radar scatterers and lower signal-to-noise ratios (SNRs), leading to this region often being filtered out and ignored in radar analyses of supercells and tornadogenesis. This has led to most discussion of the second stage of supercell tornadogenesis, the formation of vertical vorticity near the ground, to focus on just the outflow region of supercells (Davies-Jones and Brooks, 1993; Wurman et al., 2007; Markowski et al., 2012; Marquis et al., 2012; Kosiba et al., 2013; Dahl et al., 2014; Parker and Dahl, 2015; Fischer and Dahl, 2022).

1.2 Vorticity Worms in Supercell Inflow

This focus on downdraft-generated vertical vorticity was recently challenged by Markowski (2020, 2024), who simulated a tornadic supercell in an environment with a turbulent planetary boundary layer, finding that the formation and intensity of tornado-like vortices (TLVs) is very sensitive to the realization of PBL turbulence and location of the storm-initiating warm bubble. Furthermore, Markowski (2024) found that tornadogenesis was facilitated by vertical vorticity in the storm’s inflow, rather than the outflow. This inflow vorticity is associated with turbulent coherent structures that form in the surface-layer of a sheared PBL (Robinson, 1991; Khanna and Brasseur, 1998; Kim

and Adrian, 1999; Young et al., 2002; Hutchins and Marusic, 2007; Hutchins et al., 2012). Friction near the ground leads to rapidly increasing wind speed with height, generating horizontal vorticity cross-wise to the mean ASL flow. These horizontal vortex lines are then tilted by surface-layer updraft/downdrafts, which are typically associated with the legs of hairpin vortices that are organized into rolls (Adrian, 2007; Bernard, 2011; Davidson, 2015; Watanabe et al., 2019), leading to alternating streaks of positive and negative vertical vorticity. Because this process tends to be confined to the surface-layer, this large magnitude vertical vorticity is confined to a vertical depth of just a few hundred meters above ground level (AGL). It is also important to note that substantial horizontal vorticity is still present in the ASL under these conditions as the result of friction, which has been investigated in the context of supercells and tornadogenesis (Schenkman et al., 2014; Roberts et al., 2016; Roberts and Xue, 2017). The tilting of this horizontal vorticity near the surface can generate sufficient vertical vorticity near the ground and contribute to tornado formation if it is abrupt enough, in contrast to the tilting of environmental horizontal vorticity that contributes to the formation of rotation in the mesocyclone (step 1 of the Fischer et al. (2024) model). These streaks of vertical vorticity are colloquially referred to as “vorticity worms” due to their elongated and bending nature (Gebauer et al., 2026). In the Markowski (2024) simulations, when these vorticity worms passed under the low-level mesocyclone of a supercell and interacted with gust front boundaries, their associated vertical vorticity was rapidly stretched, facilitating the tornadogenesis process. Markowski (2024) also notes that baroclinic vorticity generation does not act as a significant source of circulation for the tornado in his simulations. To confirm the validity of these findings and numerical model set up, Markowski (2024) ran an additional simulation using the same environment and employing “20th century” style parameters, such as a free-slip lower boundary condition and removing the boundary layer spin up. With this model

set up, a tornado still formed, and the baroclinic generation of vorticity in the storm’s downdraft acted as the dominant source of near ground rotation. He also found the cold pool to be 2-3 K colder than when a turbulent boundary layer was included.

The implications of this study are potentially immense, as few studies have commented on supercell inflow being a significant source of near-ground vertical vorticity for tornadogenesis. It also presents potential new environmental conditions to consider when forecasting tornadoes, including the near ground wind speed and surface heat flux in a storm’s environment. Because vorticity worms first form from frictionally generated vertical wind shear near the ground, one would expect higher wind speeds to promote large magnitudes of vorticity. Furthermore, Markowski et al. (2003), Pounds et al. (2025), and Saba et al. (2026) all found that near-surface ground-relative wind speeds are larger in tornadic environments than non-tornadic, perhaps connecting the formation of vorticity worms to tornadogenesis. However, questions remain about whether vorticity worms exist in real supercells and act as focal points for tornadogenesis.

Gebauer et al. (2026) discusses Doppler lidar observations of near-ground supercell inflow, finding multiple instances of vorticity worms near pre- and non-tornadic supercells. They used inferred vertical vorticity from the azimuthal shear of radial velocity

$$\zeta_{inf} = \frac{1}{r} \frac{\partial V_r}{\partial \theta}, \quad (1.1)$$

finding ζ_{inf} to be a sufficient parameter for identifying vorticity worms and assessing their associated stretching potential by calculating Doppler circulation within a specified radius. They do note, however, that ζ_{inf} is very sensitive to the mean wind direction in the PBL. This is because ζ_{inf} accounts for just half of the vertical vorticity term, excluding the shear of the azimuthal shear in the radial direction, which can-

not be measured with a single Doppler lidar and radar. Additionally, Gebauer et al. (2026) found that vorticity worms amplified while under the mesocyclone and near storm-scale boundaries, similar to the Markowski (2024) simulations. They also note that vorticity worms appear pseudorandom in their observations and span a range of spatial scales (≈ 50 -500 wide). This is particularly interesting given the findings of Parker (2023), which investigated the organization of surface-layer vertical vorticity needed for tornadogenesis. Parker (2023) found that while randomly organized vertical vorticity can be sufficient for tornadogenesis, “coherent vorticity sources that produce longer-wavelength structures are likely to be the most supportive of tornadogenesis” (p. 1). These studies highlight a need to characterize the spatial scale of vorticity near the ground in supercells, both in the inflow and outflow, to identify which structures are most conducive of tornadogenesis.

1.3 10 May 2010 Oklahoma Tornado Outbreak

To investigate the nature of near-ground vertical vorticity in a tornadic supercell, this study primarily investigates the Central Oklahoma tornado outbreak on 10 May 2010, which was observed by OU-PRIME (Palmer et al., 2011; Griffin et al., 2017), which was a high-resolution polarimetric C-band radar operated by the University of Oklahoma. This outbreak featured 56 tornadoes in Oklahoma, including 2 EF4s, 4 EF3s, 7 EF2s, and 21 EF1s (National Centers for Environmental Information, 2026). Both EF4 tornadoes occurred in the southeast Oklahoma City metro area in close proximity to OU-PRIME. The first EF4 touched down in Moore, Oklahoma, north-northwest of OU-PRIME, and tracked north east into Choctaw and Harrah, Oklahoma. Given the inflow winds were south-southeasterly, the OU-PRIME beam was well aligned, allowing the use of inferred vertical vorticity from the azimuthal shear of radial velocity

in this case. OU-PRIME operated with a 0.45° half-power beamwidth, and the volume coverage patterns included a 0.25° elevation angle, providing a rare opportunity to characterize near-ground (< 100 m AGL) vertical vorticity in all regions of the Moore-Choctaw supercell during tornadogenesis. This supercell also passed through the center of a dual-Doppler lobe between OU-PRIME and the KTLX WSR-88D radar (see Fig. 3.2), which allows for the retrieval of three-dimensional winds and provides full-storm context. Palmer et al. (2011) notes the presence of many interesting features in this case, including the existence of vortices along the RFGF of the Moore-Choctaw supercell that increased in size with height. This combined with vorticity worms in the environmental inflow provides an opportunity to analyze the amplification of inflow vorticity along storm-scale boundaries and under the low-level mesocyclone in an observed tornadic supercell for the first time. Using this data set, the following questions will be investigated:

1. How does the wavelength of near-ground vertical vorticity compare across different regions of the Moore, Oklahoma supercell?
2. How did the wavelength of near-ground vorticity evolve over time and correlate with inflow/outflow wind speeds?
3. How does near-ground vertical vorticity and circulation compare between the inflow and outflow?

The velocity retrievals in the inflow of the Moore supercell are a rare dataset as many radar studies have too few scatterers in clear-air to properly observe the near-surface inflow. The storm's proximity to OU-PRIME also provides data from multiple elevation angles in the surface layer in the inflow, which will allow for detailed analysis of vorticity worms. Given the findings of Parker (2023), analyzing the wavelength of surface-layer vertical vorticity across all storm regions can provide insight into which spatial scales

of vertical vorticity are more supportive of tornadogenesis. Viewing the wavelength over time and comparing to storm-scale phenomena will explain how processes like inflow intensification or surges of the rear-flank downdraft (RFD) influence vorticity worms. Lastly, the inflow is hypothesized to provide sufficient vertical vorticity for tornadogenesis. Comparing available vertical vorticity and circulation between the inflow and outflow will provide insight into this hypothesis.

Chapter 2

Background

Coherent streaks of vorticity in the surface layer of the PBL (“vorticity worms”) have often been overlooked in studies of tornadoes (as discussed in Chap. 1), and thus, are not well understood in the severe storms community. To understand vorticity worms and their formation, one must first understand the different regions of the PBL.

2.1 Planetary Boundary Layer Structure

During the day, a convective boundary layer (CBL) forms, forced by positive heat fluxes from the ground. Nearest the surface in the lowest 10% of the CBL is the atmospheric surface layer (ASL). In contrast to the rest of the CBL above, the ASL is most commonly denoted by a logarithmic wind profile that increases with height (Fig. 2.2) due to surface friction. Under convective/stable conditions, the ASL has a rapid decrease/increase in potential temperature with height. Above the ASL during the day is the convective mixed-layer, spanning the rest of the depth of the CBL, then the entrainment zone above that, which denotes a transition zone from the PBL to the free troposphere (Fig. 2.1).

As previously mentioned and shown in Fig. 2.2, the ASL typically consists of rapid wind speed and potential temperature changes with height. The convective mixed-layer, which makes up the majority of the PBL above the ASL, consists of approximately constant wind speed and potential temperature with height. This is

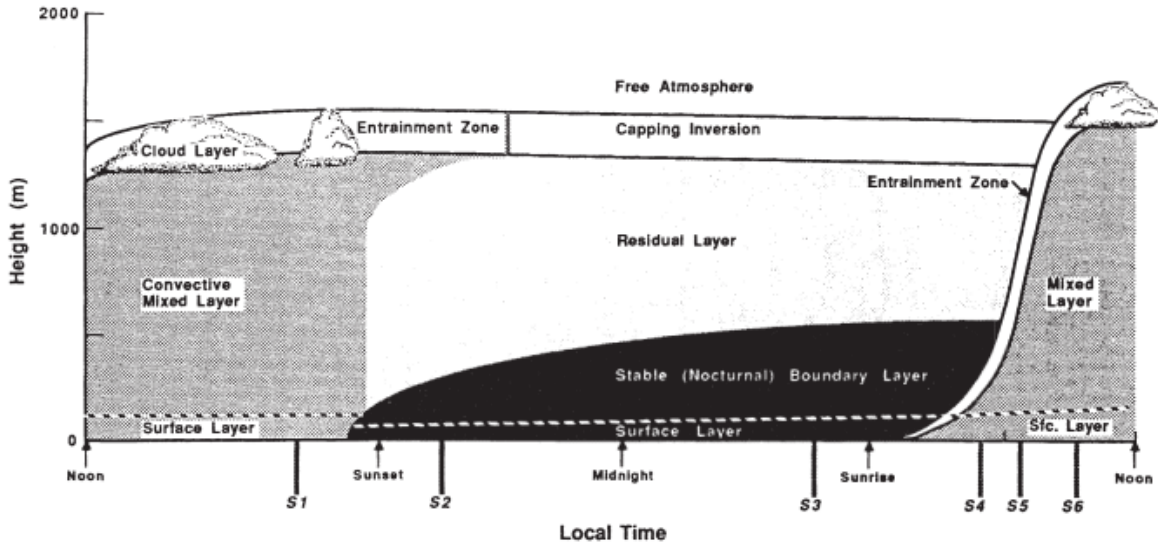


Figure 2.1: Diagram of the diurnal cycle of the atmospheric boundary layer from Figure 1.7 in Stull (1988).

due to the turbulence of the mixed-layer, which is typically convectively driven during the day. In the entrainment zone, there is a potential temperature inversion, which denotes the top of the CBL. The wind speed profile also increases from sub-geostrophic to geostrophic within the entrainment zone.

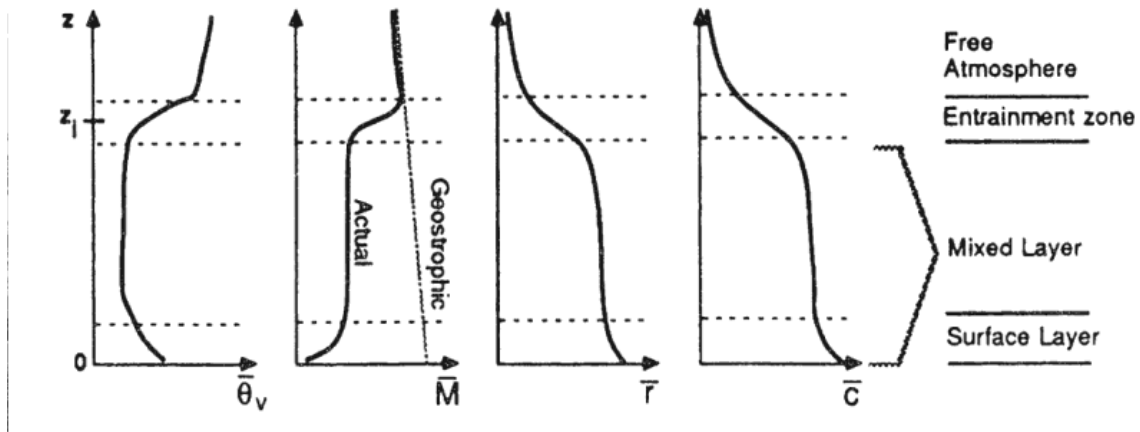


Figure 2.2: From left to right, typical virtual potential temperature ($\bar{\theta}_v$), wind speed ($\bar{M}$), water vapor mixing ratio ($\bar{r}$), and pollutant concentration ($\bar{c}$) profiles in the convective boundary layer from Figure 1.9 in Stull (1988).

When the sun sets, a stable boundary layer (SBL) forms as the surface heat fluxes transition from positive to negative (Fig. 2.1). Above the SBL is the residual layer, where turbulent mixing from the previous day's CBL is still present. Lastly, a capping inversion remains where the entrainment zone once was atop the residual layer. While no longer convectively unstable, a rapid increase in the wind speed close to the ground is still present, generating large shear within the SBL (Fig. 2.3). The potential temperature now increases logarithmically in the surface layer, indicating strong static stability near the ground at night. The residual/mixed layer is still denoted by a constant potential temperature with height, then another inversion in the capping inversion. Another difference between the convective and stable boundary layer is the existence of super-geostrophic winds near the top of the stable boundary layer and bottom of the residual layer.

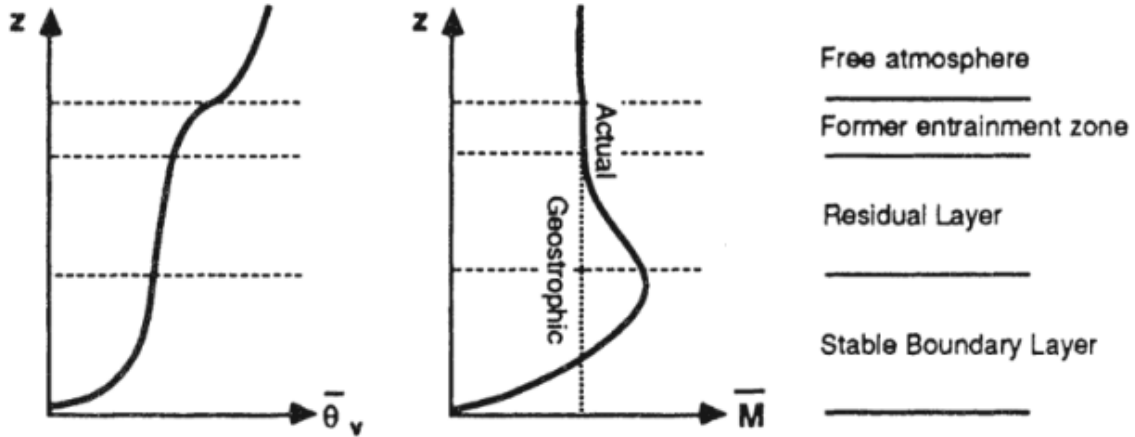


Figure 2.3: Typical virtual potential temperature (left, $\bar{\theta}_v$) and wind speed (right, $\bar{M}$) profiles in the stable boundary layer from Figure 1.11 in Stull (1988).

The super-geostrophic winds in the stable boundary layer are associated with the nocturnal low-level jet (NLLJ) (Blackadar, 1957). The NLLJ forms as the PBL becomes stable because vertical motion is generally suppressed due to the stability near the ground. Blackadar (1957) describes this diurnal phenomena as an inertial oscillation,

as the boundary layer winds transition from sub- to super-geostrophic as the SBL forms. As the winds atop the SBL accelerate beyond the geostrophic wind, very strong low-level shear forms, generating turbulence. While the mechanically generated turbulence near the ground overnight may not be as strong as convectively driven turbulence during the day, the strong low-level shear generates ample horizontal vorticity in the surface layer.

2.2 Turbulent Coherent Structures

As noted in Chap. 1, Markowski (2024) found streaks of alternating positive and negative vertical vorticity at the lowest model level in his supercell simulations that included a turbulent boundary layer. The Markowski (2020, 2024) simulations excited turbulence by including random temperature perturbations (± 0.25 K) in an environment with a mixed layer in the lowest ≈ 1 km AGL, then running the model for 12 hours to allow for the formation of a steady-state turbulent boundary layer. It is important to note that the surface heat flux in his simulations was zero, meaning the boundary layer is neutrally stratified. Markowski (2024) describes the streaks of vertical vorticity as shear driven turbulent coherent structures, which have been extensively studied in the boundary layer meteorology and turbulence communities (Robinson, 1991; Khanna and Brasseur, 1998; Young et al., 2002). The streaks also manifest as streamwise velocity perturbations, or alternating high- and low-speed streaks) in the ASL, which are associated with the positive and negative vertical vorticity in the Markowski (2024) simulations. While these structures appear quite straight and coherent in Markowski (2024), they tend to bend more in the atmospheric boundary layer due to small-scale perturbations and changing land surfaces, leading to these structures being coined “vorticity worms” by Gebauer et al. (2026).

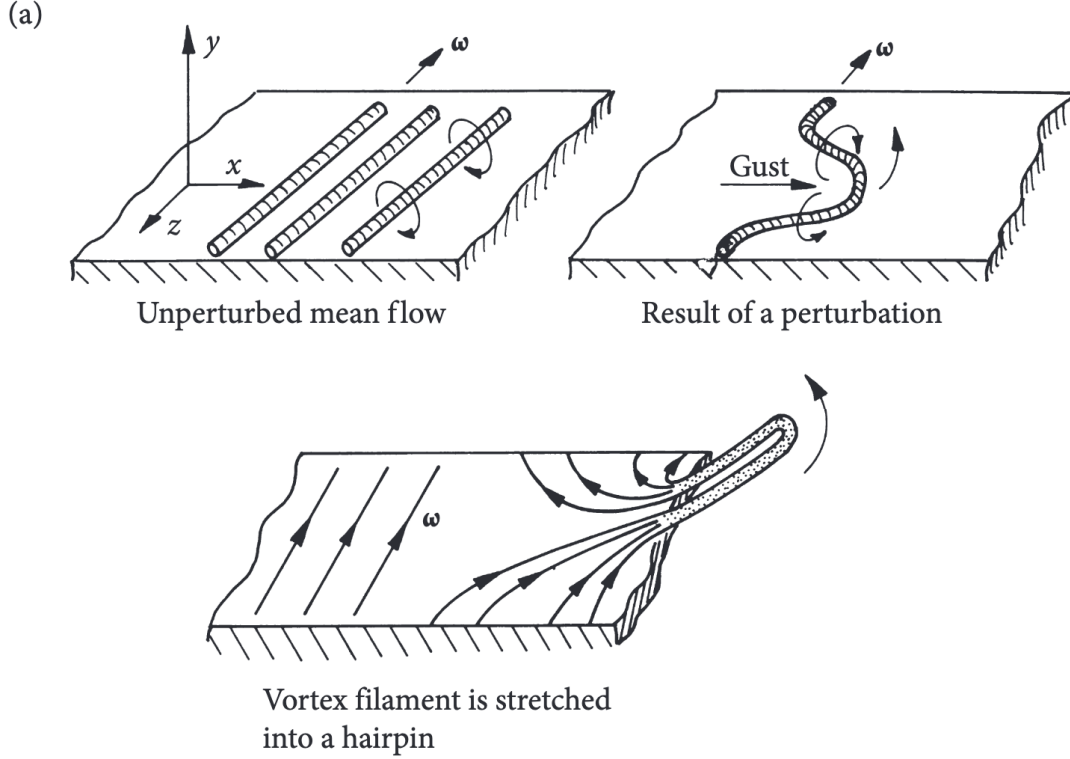


Figure 2.4: The formation of a hairpin vortex from Fig. 4.3a in Davidson (2015).

Vorticity worms form distinctly from the more commonly known horizontal convective rolls (HCRs), which are convectively driven streaks of updrafts and downdrafts that span the depth of the mixed layer in sheared convective boundary layers (Khanna and Brasseur, 1998; Sullivan et al., 1998; Salesky et al., 2017). Vorticity worms are often referred to as “superstructures” in the boundary layer and engineering communities and are found to be constrained to the surface layer (Kim and Adrian, 1999; Hutchins and Marusic, 2007; Hutchins et al., 2012). These superstructures can have stream-wise length scales that are 20 times the depth of the surface layer (Kim and Adrian, 1999; Hutchins and Marusic, 2007) and are often the conglomeration of multiple hairpin vortices (Watanabe et al., 2019). Hairpin vortices are widely studied structures in the surface layer that have parallel counter-rotating vortices and are connected by a

horseshoe-shaped arch (Adrian, 2007; Bernard, 2011; Davidson, 2015). Hairpin vortices are thought to form in multiple fashions, the first being by the bending of environmental vortex lines by streamwise velocity perturbations (like a gust of wind), then the tilting of this bend by some vertical motion (shown in Fig. 2.4). Because the wind speed increases logarithmically with height in the surface layer, the hairpin can be stretched when tilted vertical in this manner. The second formation mechanism of hairpin formation is the “pinching” of horizontal vortex lines by PBL deep eddies, such as horizontal convective rolls (Fig. 2.5c). Hairpin vortices are commonly oriented where their counter-rotating vortices are parallel to the mean wind in the ASL (Fig. 2.5b). As these counter-rotating vortices are aligned into coherent streaks, so is the vertical vorticity near the ground. As these horizontal vortex lines are pinched up, they slow the horizontal wind speed and introduce streamwise velocity perturbations, known as low-speed streaks. While hairpin vortices do not exist as clearly in the real atmosphere as in Fig. 2.4 and 2.5 due to its high Reynold’s number, the processes leading to their formation remain, allowing for the generation of vorticity worms in the real atmosphere.

While significant vertical vorticity is present in vorticity worms, it is also important to note the presence of horizontal vorticity from the large vertical wind shear in the ASL. Schenkman et al. (2014) simulated a supercell with a semi-slip lower boundary condition to model friction near the ground, finding that the tilting of horizontal vorticity generated by friction near the ground was the dominant source of vertical vorticity for tornadogenesis. This finding is especially interesting, as their simulations did not spin up a turbulent boundary layer like Markowski (2024), meaning the horizontal vorticity near the ground is purely the result of friction and there is no pre-existing vertical vorticity in the environment, yet a strong tornado still formed.

While one might expect more efficient tilting of horizontal vorticity near the ground

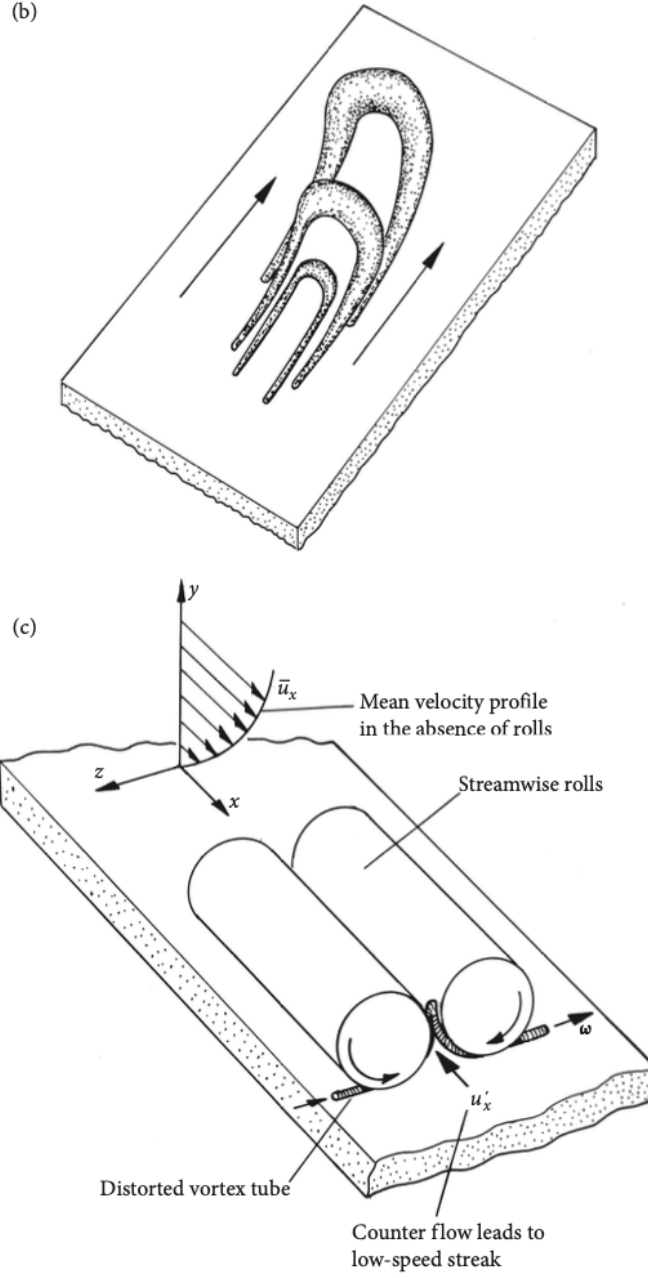


Figure 2.5: Packet of hairpin vortices (superstructure, b) and the deformation of a horizontal vortex by PBL deep eddies from Fig. 4.3 in Davidson (2015).

in the CBL due to larger vertical motion, Greene and Salesky (2024) found that stream-wise velocity perturbations (i.e., high- and low-speed streaks) can still exist under weakly stable stratification, but these coherent structures become smaller and less pro-

nounced as stability increases. This finding is consistent with those of Gebauer et al. (2026), which observed vorticity worms in both the inflow (typically unstable or neutrally stratified) and the outflow (typically stably stratified) of supercells. As previously mentioned, the large vertical shear near the ground in SBLs generates mechanically driven turbulence and streamwise velocity perturbations, which promotes the existence of vorticity worms under these weakly stable conditions (Greene and Salesky, 2024).

2.3 The M24 Tornadogenesis Mechanism

While most studies of supercell tornadogenesis have argued that vertical vorticity near the ground is generated baroclinically, Markowski (2024) (M24) found vorticity worms, which are produced by PBL turbulence, can facilitate tornadogenesis in simulated supercells. Furthermore, he found that baroclinic contributions to material circuits were negligible, and that the subgrid-scale turbulence scheme was a major contributor to the circulation. A schematic of the important features leading to tornadogenesis from Markowski (2024) is presented in Fig. 2.6. On the bottom, the ASL is shown, with the vortex lines generated by frictionally generated wind shear shown by blue arrows. The yellow tubes can be thought of as the counter-rotating “legs” of hairpin vortices, with alternating updraft and downdraft pairs. These variations in vertical velocity in the ASL lead to the alternating streaks of positive and negative vertical vorticity, which is denoted by the bending of the blue vortex lines. Slow-speed streaks are shown in the updraft regions of the hairpin vortices while fast speed streaks are seen in their downdrafts. On the top, the supercell is seen, with clouds denoted by the opaque gray and the storm’s gust front labeled by the gray fronts on the ground. The progression of vortex lines from the ASL are shown by the blue lines, with initial vortex lines following the bending path through hairpin vortices, then later lines quickly erupting into

the vertical beneath the mesocyclone. Furthermore, the storm-relative environmental winds and the blue vortex lines in the ASL are not shown as parallel, indicating the presence of streamwise vorticity in the storm’s turbulent inflow. Ultimately, as the streamwise horizontal and vertical vorticity in the ASL passes under the mesocyclone and interacts with storm-scale boundaries, rapid tilting and stretching occurs, and the tornadogenesis process occurs.

2.4 Environmental Controls of Vorticity Worms

While the above tornadogenesis hypothesis has not yet been observed in nature, (Markowski, 2024, p. 512) notes that if “ ζ streaks actually facilitate [tornado] formation in real storms, the implications are potentially enormous, as tornado forecasting should then also consider whether environmental conditions might favor hairpin vortices or any other coherent turbulent structures”. While the presence of hairpin vortices and low-speed streaks in the ASL have been noted extensively by the boundary layer and turbulence communities, the specific environmental conditions that support or suppress their development in the atmosphere, especially in storm environments, remain relatively unclear. Markowski (2024) hypothesizes that strong ground-relative winds and weak surface heat fluxes are most supportive of vorticity worms and notes how in Markowski et al. (2003), tornadic environments indeed had stronger ground-relative winds near the surface.

Frost et al. (2025) performed idealized LES of the CBL with a temporally changing surface heat flux (Q_0) and a range of constant geostrophic wind forcing (U_g). The use of a temporally changing Q_0 is motivated by phenomena such as the passing of clouds, which generate near instantaneous changes in the surface heat fluxes. Using these simulations, we present domain-averaged horizontal and vertical vorticity magnitude in

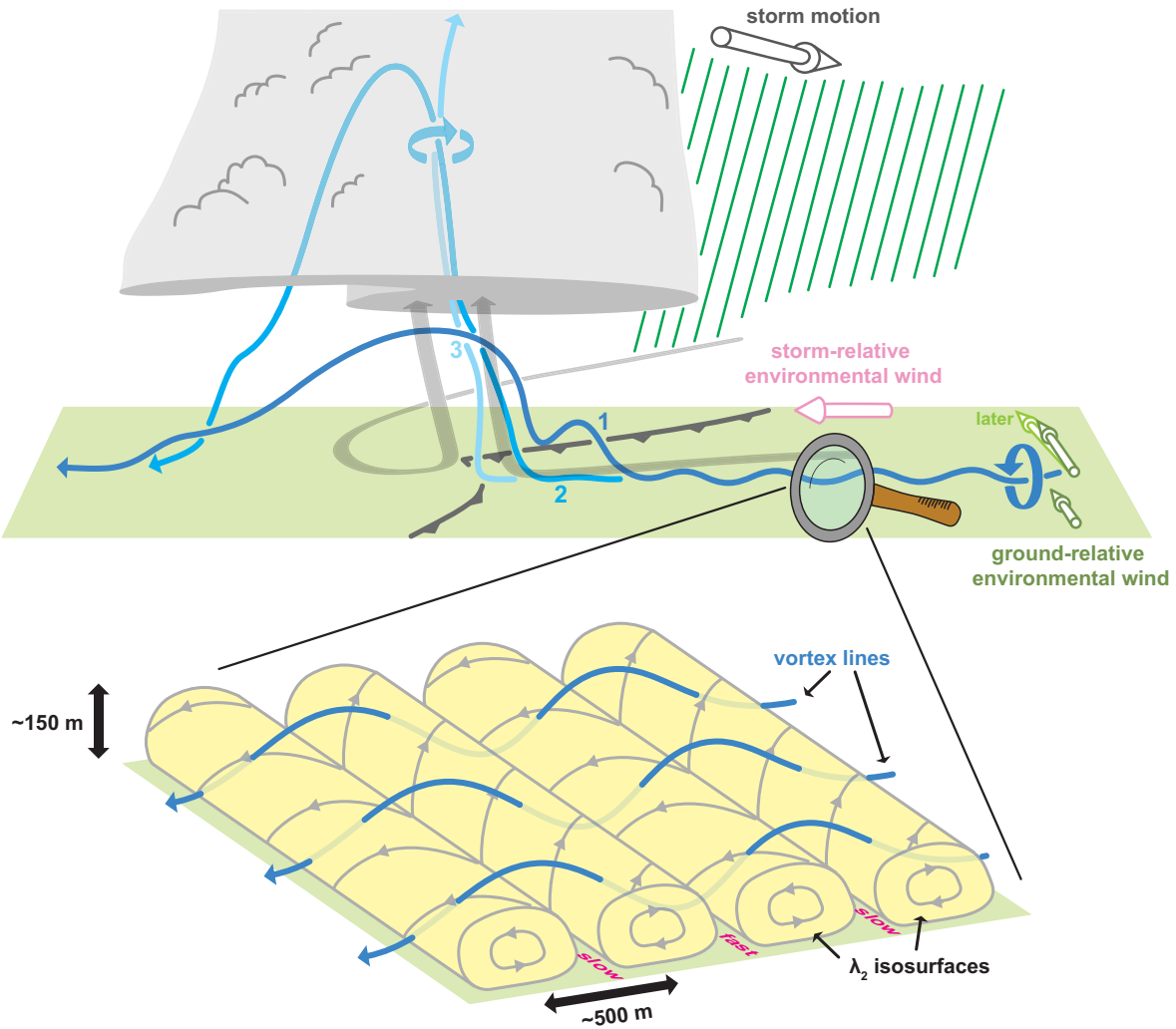


Figure 2.6: Figure 25 from Markowski (2024): Schematic summarizing the key aspects of the STORM9 simulation associated with tornadogenesis. The vortex lines (shades of blue) labeled “1,” “2,” and “3” can represent either a single vortex line at multiple times through a parcel approaching the developing TLV, a la Fig. 13, or different vortex lines at a single time drawn through locations increasing upstream of the developing TLV. The broad gray arrows indicate (storm-relative) trajectories feeding the TLV from the environment and from the precipitation and outflow (precipitation/outflow trajectories enter the developing TLV only once the flow associated with the intensifying z maximum becomes highly “axisymmetrized”). In the close-up of the environmental surface layer (bottom of figure), the gray streamlines indicate ground-relative perturbation winds associated with surface-layer eddies. The locations of alternating surface streaks of fast and slow wind speed are also indicated.

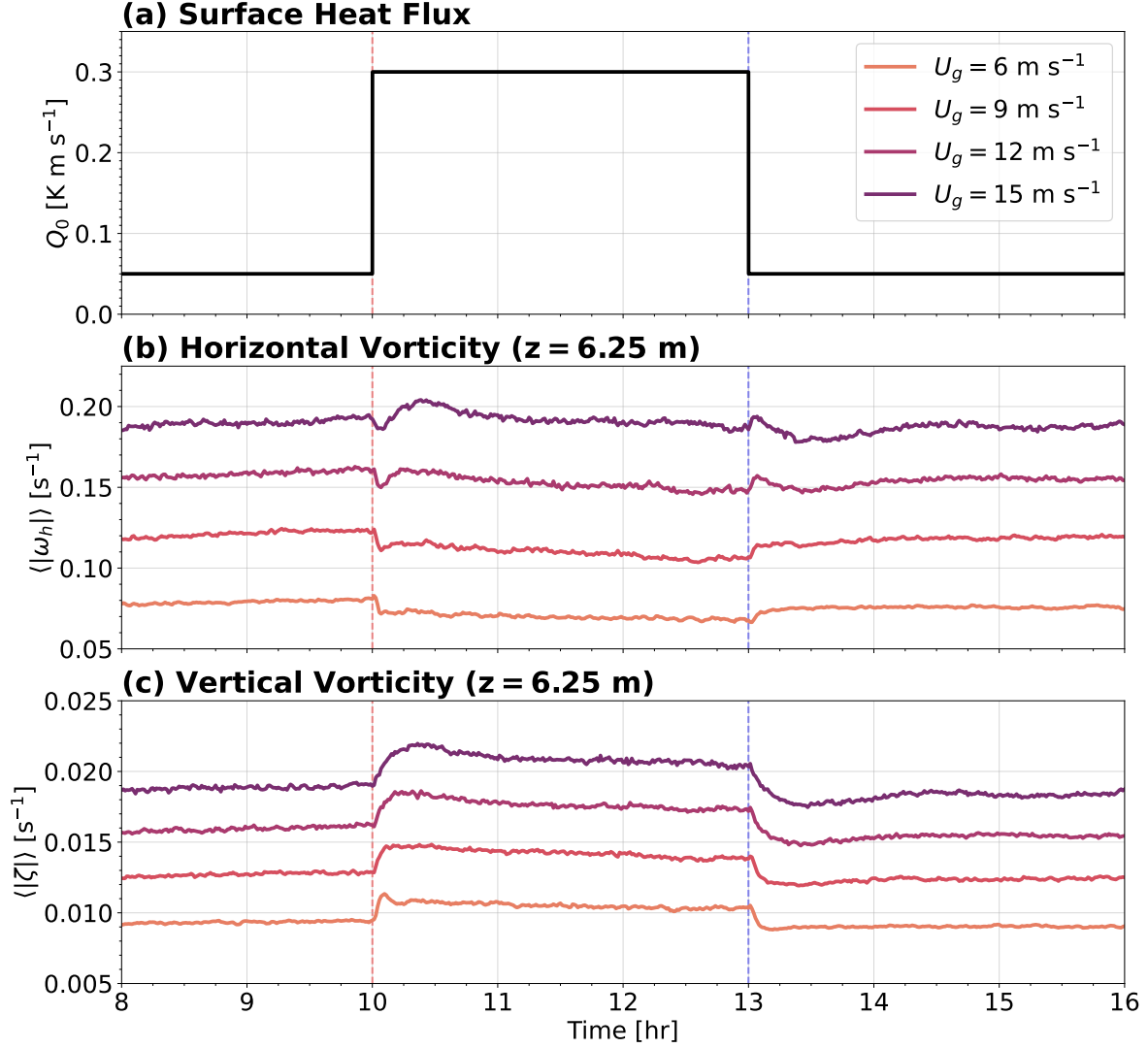


Figure 2.7: Imposed surface heat flux (a), and domain averaged horizontal (b) and vertical (c) vorticity magnitude at $z = 6.25$ m AGL for four different imposed geostrophic winds from the LES in Frost et al. (2025).

Fig. 2.7 at the lowest model grid point ($z=6.25$ m). It is shown that both horizontal and vertical vorticity magnitude increases with the geostrophic wind speed. As mentioned previously, this is because larger pressure gradient forcing (and thus, geostrophic winds) generates larger vertical wind shear in the ASL (see Fig. 4d in Frost et al. (2025)), and thus, larger cross-wise horizontal vorticity. In addition, when the surface heat flux is increased, the vertical vorticity magnitude increases with a lag on the order of ≈ 10 -30

minutes. It also decreases back to its original magnitude following the surface heat flux decrease later in the simulations. This represents the tilting of horizontal vortex lines in the ASL by convective thermals, which become more intense under larger surface heat fluxes. While this analysis shows that larger pressure gradient forcing and surface heat fluxes support larger magnitudes of both horizontal and vertical vorticity in the ASL, it is unclear how these conditions influence the coherence and “streakiness” of vorticity worms, indicating a need for further investigations of surface layer vorticity in simulations and observations.

Chapter 3

Data and Methods

3.1 OU-PRIME

OU-PRIME was a high-resolution, 5.44-cm wavelength radar operated by the Advanced Radar Research Center at the University of Oklahoma (described in full by Palmer et al. (2011)). The use of C-band while still using an 8.5-m diameter dish gave OU-PRIME at 0.45° half-power beamwidth, providing very high resolution data for severe weather research (Kumjian, 2011; Picca and Ryzhkov, 2012; Bodine et al., 2014). OU-PRIME also operated with 1000 kW peak power and 50 dB gain (Palmer et al., 2011), providing higher sensitivity to weak echoes than the traditional Weather Surveillance Radars (WSR-88D) used in the NEXRAD network. During The 10 May 2010 central Oklahoma tornado outbreak featured 56 tornadoes in Oklahoma, including two supercells in the southeast Oklahoma City metro, which each produced EF4 tornadoes at close proximity to OU-PRIME (National Centers for Environmental Information, 2026). The first occurred from 22:20 – 22:51 UTC and tracked from Moore to Harrah, Oklahoma (referred to as storm/tornado I1 herein). The second supercell (J1 herein) approached OU-PRIME from the south and tornadogenesis occurred extremely close to OU-PRIME. The FFD precipitation from storm J1 also merged with the mesocyclone of storm J1 after genesis of the I1 tornado. Due to the current study’s focus on analyzing all regions of the storm during tornadogenesis, storm I1 is the primary focus

of the current study.

This study analyzes the 8 volume scans performed by OU-PRIME between 22:09–22:28 UTC to focus on the tornadogenesis and tornado intensification period. During this period, storm I1 was located north-northwest of the radar as tornadogenesis occurred. This positioning, combined with a predominately south-southeasterly inflow winds, allows for decreased error in estimating vertical vorticity by using the azimuthal shear of radial velocity (ζ_{inf}). OU-PRIME operated with sector scans at 0.25° , 1.0° , 2.0° , 3.0° , 4.0° , 5.0° , 6.5° , and 9.0° elevation angles. Volume updates were every ≈ 2.5 minutes and range resolution was 125 m. During this analysis period, the mesocyclone of storm I1 was at a range of 15–26 km from OU-PRIME. This yields a beam-height of 80–150 m AGL and a cross beam distance of 135–225 m in the mesocyclone for the 0.25° sweeps. This means the inflow region, which is closer to the radar for all times, was observed 50–100 m AGL and with a cross beam distance of 50–100 m. Data are quality controlled as in Griffin et al. (2017); aliased radial velocities are manually unfolded and ground clutter is manually identified as stationary, high power targets with near-zero radial velocity and removed. In addition, data where $\text{SNR} < 0$ dB are excluded from the analysis. These high-resolution observations of supercell inflow are rare and allow us to characterize the nature of vertical vorticity worms in very fine detail using OU-PRIME.

3.2 Pseudo Vertical Vorticity

As in Gebauer et al. (2026), vertical vorticity is inferred from the azimuthal shear of radial velocity (ζ_{inf} , Eq. 1.1). It is important to note that ζ_{inf} only captures half of the full vertical vorticity term, as it excludes the azimuthal shear of the azimuthal velocity $\left(\frac{1}{r} \frac{\partial(r u_\theta)}{\partial \theta}\right)$ because u_θ cannot be measured by Doppler radar/lidar. However, when the

radar/lidar beam is well aligned with the mean wind, the vertical vorticity associated with structures like vorticity worms is well resolved by ζ_{inf} . The ζ_{inf} calculation is done in radar coordinates and is performed with centered, finite differencing of the radial velocity field, and forwards or backwards differencing on the edges to maintain the same shape as the radial velocity field. For the wavelet analysis, ζ_{inf} is then gridded with a Barnes analysis before the wavelet transform is calculated (full described in Chap. 3.3). For some plots, radial velocity is smoothed with a rolling median before calculating ζ_{inf} to remove noise and more easily highlight individual vorticity worms.

3.3 2D Wavelet Transform

To provide a quantitative measurement of the wavelength of inferred vertical vorticity across different regions of storm I1, we perform a two-dimensional multi-level decomposition using the Haar wavelet transform (Haar, 1910). While the Fourier transform can provide high spectral-resolution, it does not provide information on the location of signal in time or space. The wavelet transform breaks signals down into different frequency ranges, allowing for analysis of both what frequencies are present and when/where they occur. The discrete wavelet transform has been used extensively in digital image processing, as the retention of both frequency and spatial information is particularly useful for image filtering and compression (Porwik and Lisowska, 2004). An example of the discrete Haar wavelet transform applied to a signal with multiple frequencies over time is presented in Fig. 3.1. Note how the temporal resolution is very poor at large frequencies, then becomes more fine at higher frequencies. When applied in space, the spatial resolution of the discrete wavelet transform is finer for smaller wavelengths and coarser for larger wavelengths. During the first period, when the dominant frequency is $f = 5$ Hz, most power appears around this range, while some

power exists at other frequencies due to the random perturbations (Fig. 3.1). During the second period, most power is present around $f = 20$ Hz. During the final period, when multiple dominant frequencies are present, the largest power is seen around $f = 5$ and 10 Hz.

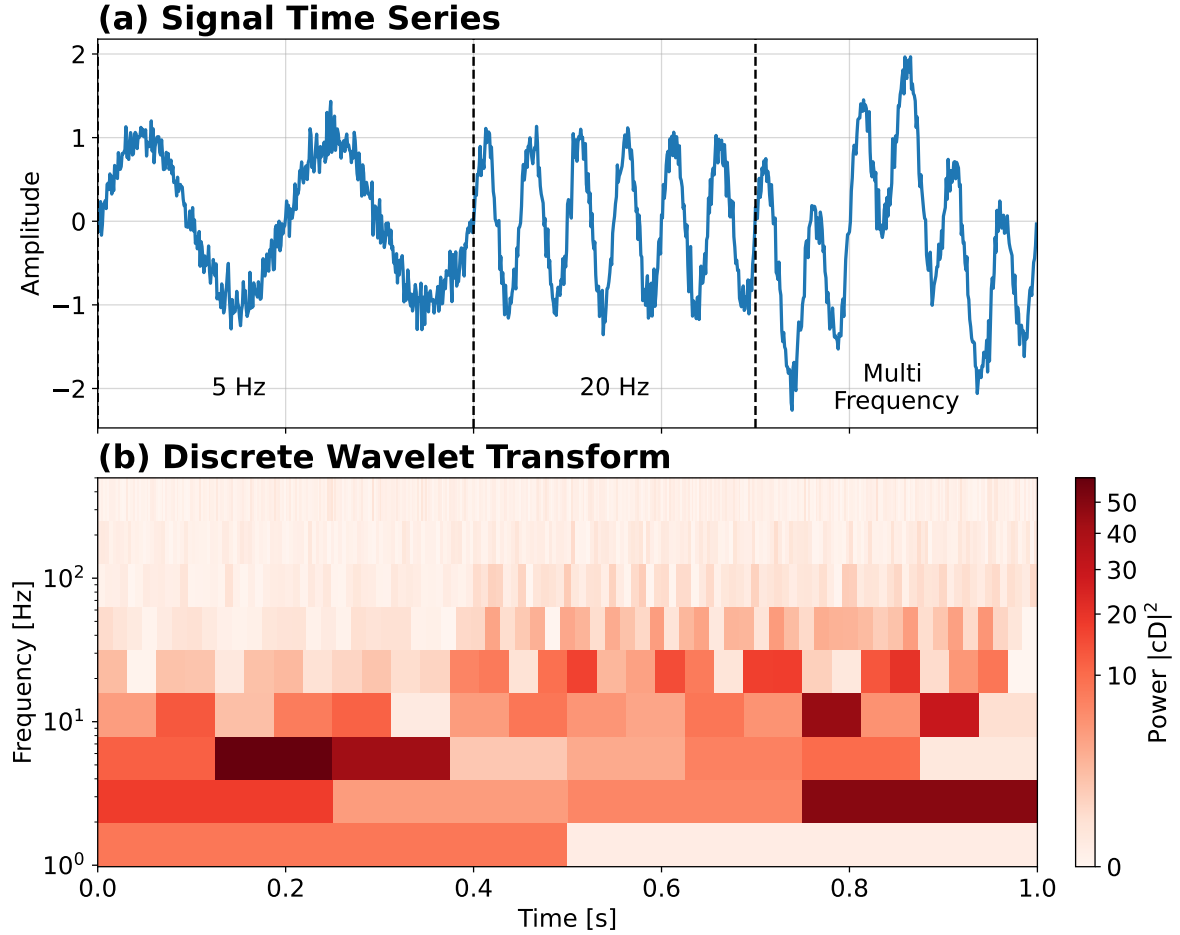


Figure 3.1: Time series of a signal with multiple frequencies over time and random perturbations (a) and an example scalogram from a discrete Haar wavelet transform (b).

Wavelet analysis has also proven useful for dissecting turbulent flow fields into their multi-scale flow components (Rinoshika and Rinoshika, 2020). Orthogonal wavelets, such as the Haar function, are particularly useful for decomposing flow fields because the coefficients at different levels are orthogonal, allowing features at different spatial

scales to be analyzed isolated from other wavelet levels (Daubechies, 1992). Wavelet analysis was also used by Liu et al. (2007) to detect tornadoes using radial velocity observations from the operation Next Generation Weather Radar (NEXRAD) network. Their use of the Haar wavelet allowed for examination of tornado velocity signatures (TVSs) at multiple spatial scales and better discrimination between non- and tornadic vortices. Ultimately, while the wavelet transform provides coarser spectral resolution than the Fourier transform, the retention of spatial information allows for the viewing of power spectral density on a map, enabling the comparison of integrated power for wavelength ranges in different storm regions. To simplify the interpretation of the wavelet analysis, we use the Barnes interpolation from Py-ART (Helmus and Collis, 2016a) to put the OU-PRIME observations on a cartesian grid, though an interpolation in the vertical dimension is not performed. This is foregone to preserve the high resolution of the OU-PRIME observations and avoid non-physical artifacts that could be introduced by a vertical interpolation. This means the radar beam-height is slightly higher in the outflow than the inflow, but the difference is on the order of ≈ 10 m. The full details of the Barnes interpolation are presented in Tab. 3.1. Before performing the wavelet transform, we remove the horizontal mean of inferred vertical vorticity from each sweep. Lastly, the two-dimensional Haar wavelet is performed using the “wavedec2” function from the PyWavelets python package (Lee et al., 2019). The Haar wavelet base function is used due to its orthogonality, which allows for simpler analysis of the individual spatial scales in the inferred vertical vorticity field. Given the Cartesian grid spacing of 100 m, we primarily view the wavelet transform at $\lambda = 200\text{--}400$ m, $400\text{--}800$ m, $800\text{--}1600$ m, and $1600\text{--}3200$ m levels. These levels span wavelengths from the typical scale of vorticity worms up to larger-scale structures like HCRs.

3.4 Partitioned Time Series

We employ partitioned time series to compare mean quantities (wavelet power, circulation, vertical vorticity flux) of the inflow, outflow, and gust front across our analysis period. First, the position of the center of the vortex (the mesocyclone before tornadogenesis and the tornado after) is identified for each OU-PRIME sweep below 3.0° . Next, the gust front is drawn using OU-PRIME radial velocities and storm-relative winds (SRW) from the DDA at the altitude closest to the radar beam height within a 4 km radius of the vortex for each OU-PRIME sweep below 3.0° . The SRW are considered in addition to the convergence of OU-PRIME radial velocities because using single-Doppler observations alone are sensitive to the look angle of the radar. In storm I1, a zero isodop (line of constant Doppler-velocity) extends north of the reflectivity hook, which could be interpreted as the gust front. However, as shown in Fig. 4.12, the SRWs are easterly, which better resembles storm-relative inflow. Furthermore, the gust front is drawn in a way that primarily attempts to separate parcels that originate in the environmental inflow from those that originated within the storm or have passed through heavy precipitation. This is done to attempt to separate vertical vorticity associated with PBL turbulence in the inflow from that formed by storm-scale processes (i.e., horizontal vorticity tilted into the vertical by storm downdrafts). Using this, the gust front partition is defined as the region within a 500 m distance from the gust front line and within the 4-km radius circle. The inflow and outflow partitions are then manually defined as the regions around the gust front and within the 4-km radius circle. When performing the partitioned time series on data from OU-PRIME, we simply use the vortex and gust front positions for that sweep. When performing the partitioned time series on the DDA, the gridded vertical levels are matched to the sweep with the closest average radar beam height in the 4 km radius circle.

3.5 Dual-Doppler Analysis

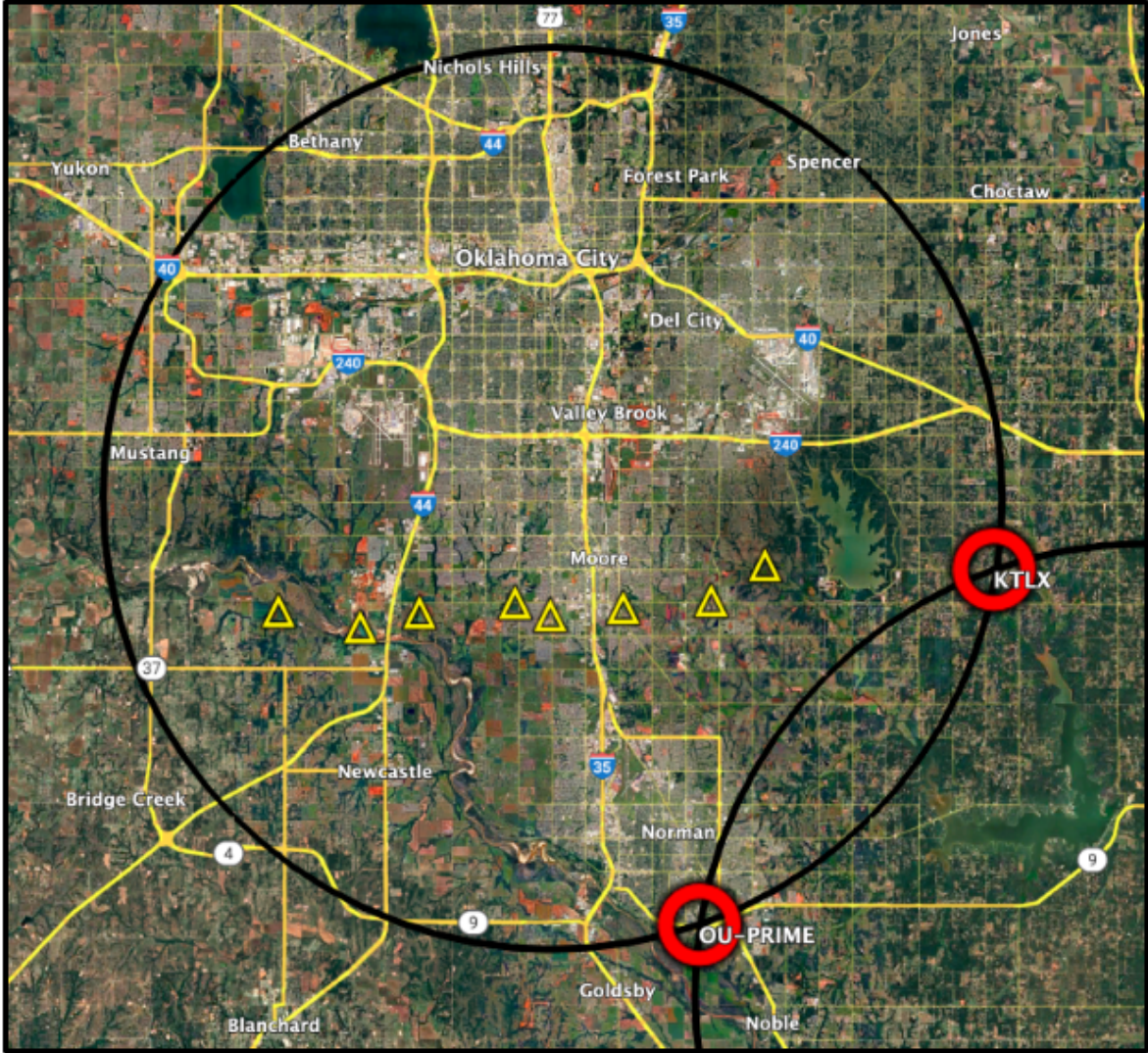


Figure 3.2: North Dual-Doppler lobe (black line) between OU-PRIME and KTLX (red circles) and subjectively analyzed track of the Moore, Oklahoma supercell during our analysis period (yellow triangles).

A dual-Doppler analysis (DDA) is performed using OU-PRIME and the KTLX WSR-88D radar to estimate the full 3-dimensional wind during tornadogenesis and assess various storm-scale attributes. We also use this DDA to analyze the statistics of vertical vorticity near the ground in storm I1, though this analysis is too coarse to

resolve most small-scale features like vorticity worms. Rather, this DDA is capable of providing broader context of the contributions of each storm region to the tornado and low-level mesocyclone circulation in storm I1. The mesocyclone track of storm I1 and a map of the north dual-Doppler lobe are presented in Fig. 3.2. Here, we see that the mesocyclone is within our DDA lobe for our entire analysis period, though it does approach the east edge of this lobe towards the end. Additionally, since KTLX is further from the mesocyclone of storm I1 and used a 0.5° lowest elevation angle, our lowest vertical level at the start of the analysis is 400 m AGL. This decreases in time as storm I1 approached KTLX.

	Radars	Grid spacing ($\Delta x, \Delta y, \Delta z$)	Vertical Interpolation	ROI Function
Wavelet Analysis	OU-PRIME	100, 100 m, N/A	None	Constant (150 m)
Dual-Doppler Analysis	OU-PRIME, KTLX	200, 200, 100 m	Linear	Constant (350 m)

Table 3.1: Parameters for the Barnes interpolation used in the wavelet and dual-Doppler analyses.

To perform the DDA, each radar was first interpolated to a horizontal cartesian grid centered on OU-PRIME with a Barnes interpolation from Py-ART (Helmus and Collis, 2016b). We then perform a linear interpolation in the vertical dimension. While a vertical interpolation is not used in the wavelet analysis, it is performed for the DDA because smaller scale features (< 200 m) cannot be resolved in the DDA, and thus, smoothing data in the vertical dimension is not detrimental to the analysis. The horizontal grid spacing is $\Delta x, \Delta y = 200$ m and the vertical grid spacing is $\Delta z = 100$ m. Next, we performed a spatially variable advection correction from Shapiro et al. (2010) to match the KTLX volume start times to OU-PRIME. Lastly, we perform the DDA using the PyDDA package (Jackson et al., 2020), which uses the 3D variational (3DVAR) technique to calculate three dimension winds from Doppler radar velocity measurements (Gao et al., 1999; Shapiro et al., 2009; Potvin et al., 2012a,b). We note

that KTLX completed volumes every ≈ 250 s and OU-PRIME every ≈ 170 s, leading to temporal differences between radars of over 1 minute for some volumes. While this de-synchronization is substantial for rapidly evolving phenomena like a supercell, we avoid using techniques like material circuits, analyzing vorticity tendency terms, and other techniques that rely on vertical gradients and focus on the mean values of vorticity and the SRWs near the ground to minimize the impact of errors in this DDA.

Chapter 4

Results

4.1 Storm I1 Evolution

OU-PRIME observed the entire tornadogenesis process and life of the subsequent EF4 tornado in storm I1 in fine detail. Many supercell signatures can be seen in Fig. 4.1, with a distinct hook echo present in reflectivity factor and a RFGF denoted by radial convergence of radial velocity east and south of the hook. The high sensitivity and angular resolution of OU-PRIME allow us to detail vorticity worms in I1's clear-air inflow using ζ_{inf} , which are present at all times in Fig. 4.1. We also note the supercell to the south (storm J1), which eventually merged with storm I1 at the end of our analysis. This storm may have played a role in intensifying the inflow of storm I1, which is shown to be extremely strong, particularly at 22:15 and 22:17 UTC (Fig. 4.1d, e), with radial velocity values exceeding 40 m s^{-1} at $z \leq 100 \text{ m AGL}$ just east of the RFGF. We also note a general increase in the magnitude of ζ_{inf} along the RFGF and inflow region over time, which could either be a physical response to the intensifying supercell, or a result of the storm moving closer to OU-PRIME, and thus the beam height being lower. Comparatively, the magnitude of ζ_{inf} is found to be much lower in the FFD of storm I1 than in the inflow and RFD. One may argue the lower ζ_{inf} in the FFD could be due to it being positioned further from OU-PRIME, leading to a larger cross beam distance and higher beam height in this region. While this is true, we note the sharp

decrease in ζ_{inf} magnitude along the edge of precipitation associated with the FFD, which indicates this signal is not purely due to radar geometry. The inflow region of I1 is shielded by the anvil (and eventually FFD precipitation) of storm J1 during our entire analysis period (not shown).

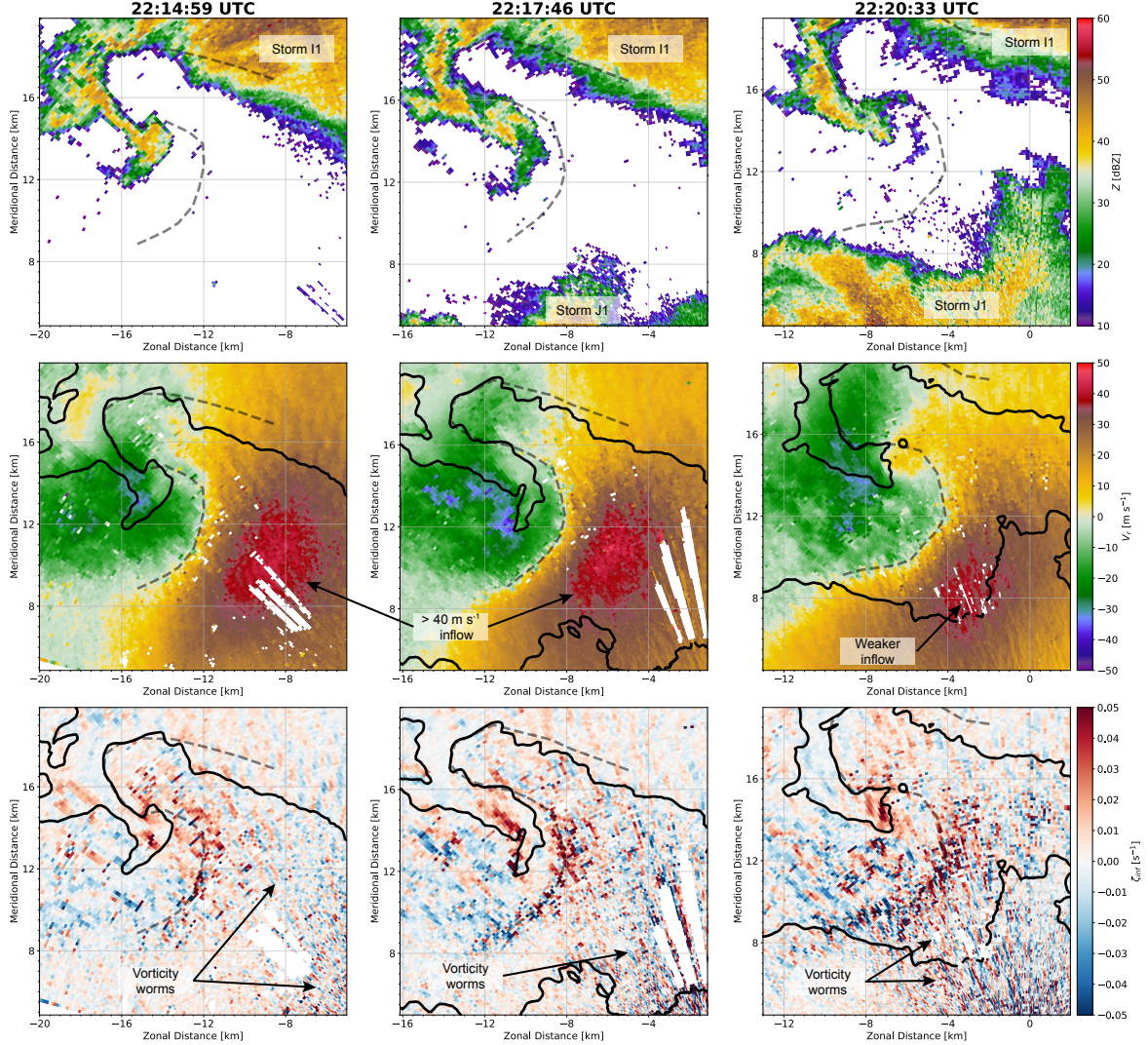


Figure 4.1: OU-PRIME 0.25° reflectivity factor (a–c), radial velocity (d–f), and inferred vertical vorticity (ζ_{inf} , g–i) from three volume scans before tornadogenesis. The solid black lines in d–i are smoothed OU-PRIME reflectivity factor contoured at $z = 20$ dBZ and the gray dashed lines are the subjectively identified gust front used for the partitioned time series.

The difference between the minimum inbound and maximum outbound radial ve-

locities at 0.25° elevation remains below tornado strength ($< 40 \text{ m s}^{-1}$) until 22:20, when it increases to over 50 m s^{-1} (Fig. 4.2a). ΔV_{max} then exceeds 70 m s^{-1} at 22:26, marking the intensification phase of the tornado. This aligns with PPIs of reflectivity and radial velocity (Fig. 4.3), where the inflow begins to surge westward into the hook. At 22:23, the hook echo appears much more narrow and cyclonically curved than the previous scan, with higher radial velocity values collocated. At 22:26, the tornado appears much more significant, with outbound radial velocities approaching 50 m s^{-1} and the mesocyclone appearing entirely occluded. The precipitation from storm J1 also begins to merge with I1, with precipitation beginning to wrap around the tornado. These observations align with the damage survey performed by NWS Norman, which lists tornadogenesis at 22:20 UTC. A significant feature in storm I1's evolution is the extreme wind speed seen in the inflow. Using the DDA between OU-PRIME and KTLX, inflow winds are found to exceed 40 m s^{-1} starting at 22:17 UTC at 200 m AGL and exceed 50 m s^{-1} above the surface layer during this period (Fig. 4.2b). Inflow winds decrease to around $30\text{--}40 \text{ m s}^{-1}$ in the surface layer at 22:23 UTC, shortly after tornadogenesis, which could be related to the occlusion process. In the RFD, wind speeds are constrained to $\approx 30 \text{ m s}^{-1}$ in the lower levels of our DDA, with peak values also occurring from 22:17–22:23 UTC.

4.2 Characterizing Near-Ground Vorticity

To characterize the vorticity worms in storm I1's inflow, we first present ζ_{inf} very near OU-PRIME in Fig. 4.4 and examine multiple elevations before tornadogenesis. Viewing ζ_{inf} at close range to the radar allows us to resolve very fine scale vorticity structures as the cross-beam distance is smaller than at further ranges ($\approx 25 \text{ m}$ in Fig. 4.4 compared to $\approx 130\text{--}220 \text{ m}$ in the mesocyclone). We also note that at this high

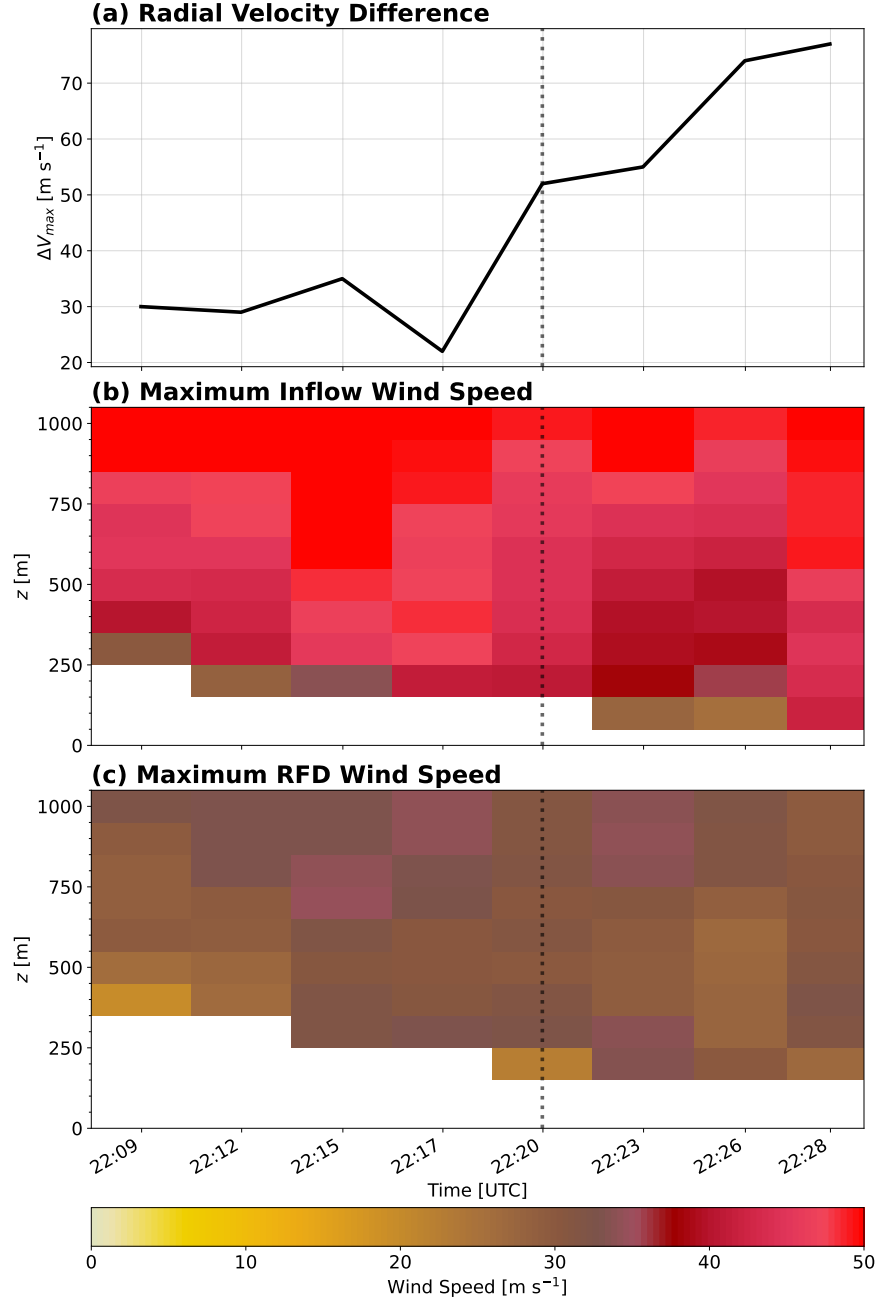


Figure 4.2: OU-PRIME 0.25° difference between the minimum and maximum radial velocity measurements within a 1-km radius of the vortex (ΔV_{max} , a), maximum inflow (b) and RFD (c) ground-relative wind speeds from DDA. The vertical gray dotted line is the approximate time of tornadogenesis.

spatial resolution, the magnitude of inferred vertical vorticity is $|\zeta_{inf}| \approx 0.1 \text{ s}^{-1}$. Using the 0.25°, 1.0°, and 4.0° sweeps, we can characterize the change in ζ_{inf} wavelength in

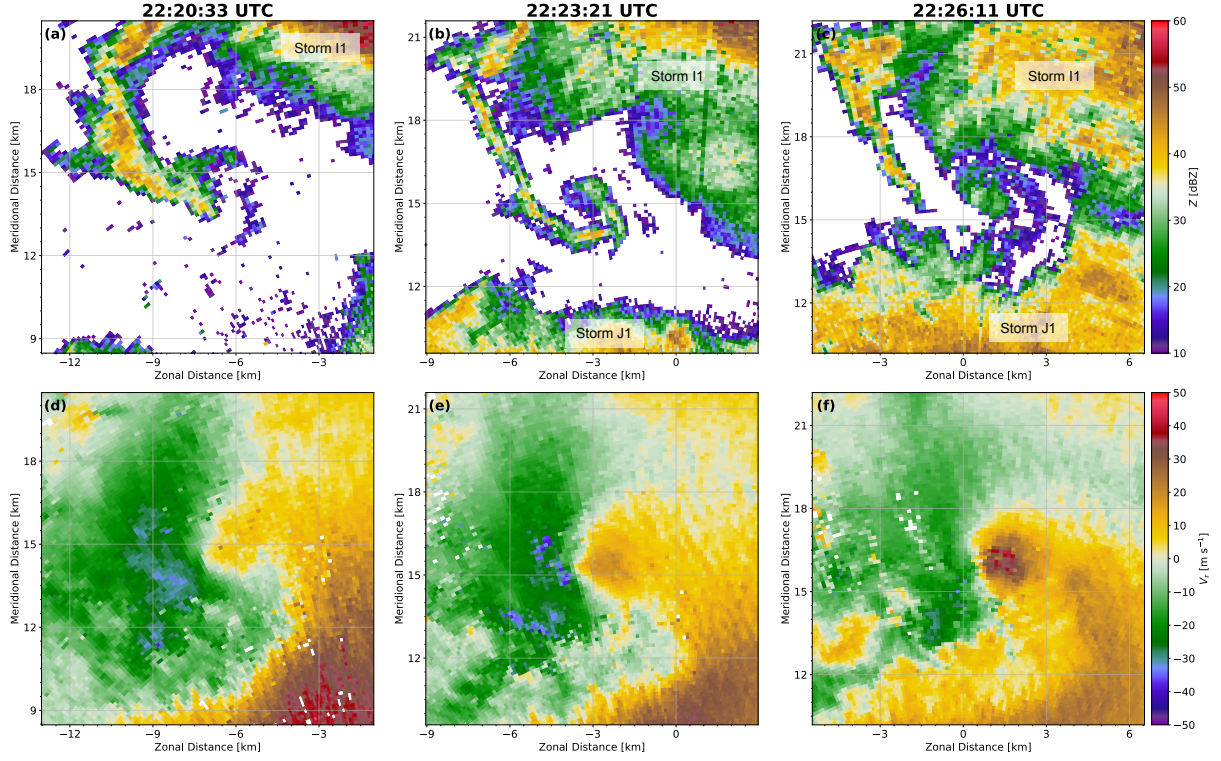


Figure 4.3: OU-PRIME 0.25° reflectivity factor (a–c), radial velocity (d–f) for three volumes around the time of tornadogenesis.

height and time by manually identifying vorticity worms that intersect the same line at a $r \approx 3$ km range. This line is drawn to be orthogonal to the mean wind in the surface layer, which is denoted with the black arrows in Fig. 4.4. At 0.25° , where the beam is 40 m AGL on average, $\lambda \approx 300$ m. This slightly increases to $\lambda \approx 400$ m on the 1.0° sweep, then up to $\lambda \geq 750$ m at 4.0° , where it becomes difficult to identify vorticity worms that intersect with the line.

The vertical structure and depth of vorticity worms remains poorly understood. Using the same technique from Fig. 4.4, we can produce a vertical profile of vorticity worm wavelength over time. This enables an examination of the depth of vorticity worms and locates the height at which structures transition to PBL-deep structures like HCRs. Modeling studies of the convective boundary layer (CBL) show that HCRs are typically over 500 m wide with wavelengths $\lambda > 1000$ m (Salesky et al., 2017;

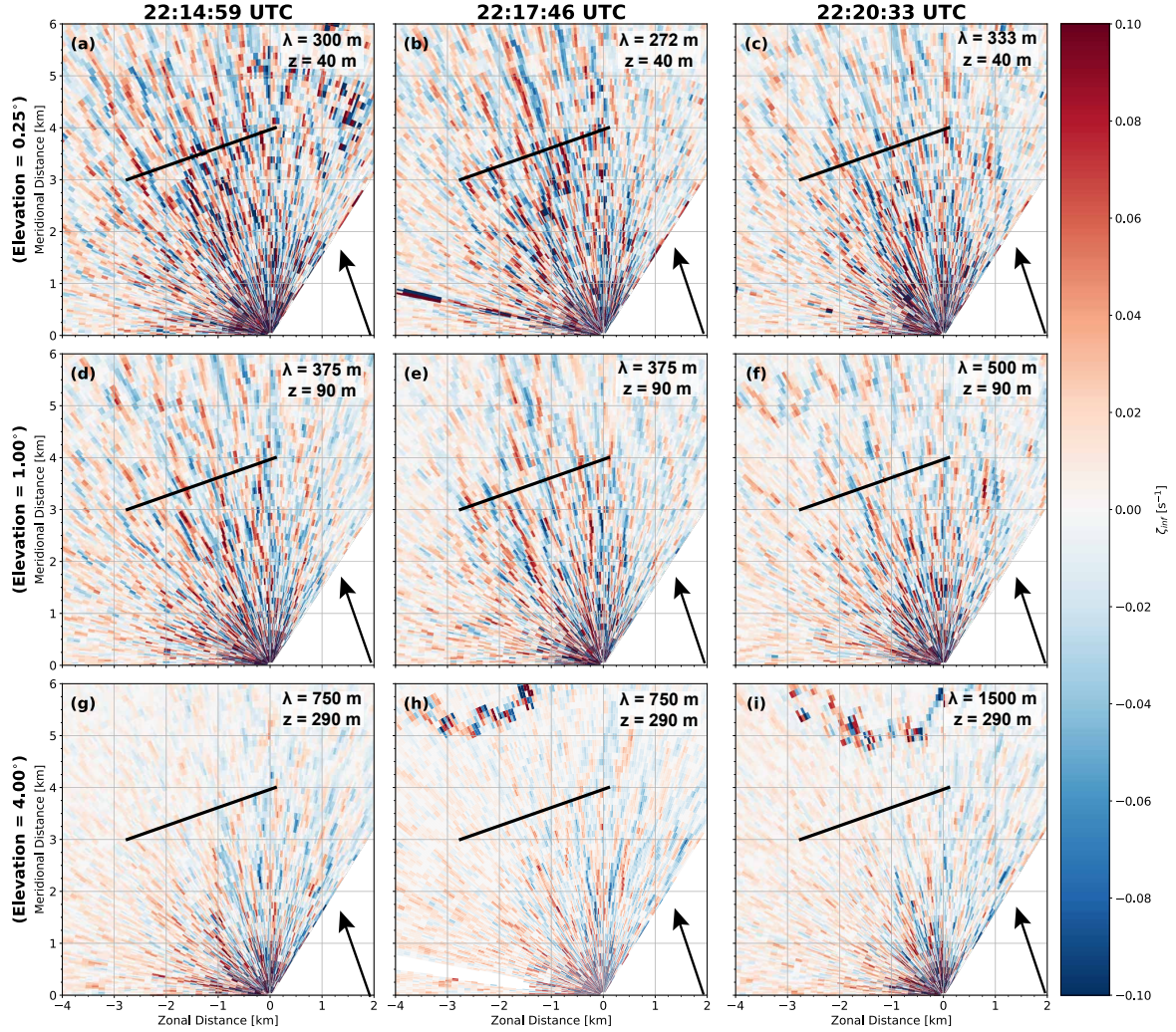


Figure 4.4: OU-PRIME ζ_{inf} from the 0.25° (a-c), 1.0° (d-f) and 4.0° (g-i) sweeps from three volumes before tornadogenesis. The black line is a 3 km wide box used to estimate the wavelength of ζ_{inf} near OU-PRIME. The black arrows represent the mean inflow wind direction and the text box is the wavelength of ζ_{inf} and average beam height along each line. Note that radial velocity is smoothed with a two gate rolling median before calculating ζ_{inf} in this figure to make it easier to see vorticity worms.

Frost et al., 2025). Before tornadogenesis (22:20), there is a substantial (≈ 100 -200 m) increase in wavelength from the 0.25° and 1.0° sweeps and the 3.0° and 4.0° sweeps (Fig. 4.5). The wavelength of vorticity worms increase with height without a sharp transition, suggesting a combination of surface-layer and HCR influence just above

the surface layer. At nearly all levels, wavelength decreases following 22:20 UTC. This is interesting, as in this case, the tornado reached its maximum intensity east of OU-PRIME. Though the area analyzed in Figs. 4.4 and 4.5 is in the inflow of storm I1 throughout the entire analysis period, the mean wind speed of this area decreased as the tornado formed and moved to its east around 22:20 (Fig. 4.2)b. Furthermore, precipitation from storm J1 moved north over OU-PRIME around 22:15, but the wavelength decrease above 50 m AGL is not until 22:23. Frost et al. (2025) found that large-scale CBL structures have a lagged response to changes in surface heat flux. The ≈ 8 minute delay between rain passing over (decreasing surface heat flux) and a decrease in the wavelength of vorticity could be related to the findings of Frost et al. (2025), or be more directly tied to changes in the ground-relative wind speed caused by storm-scale processes, such as the formation and intensification of a tornado.

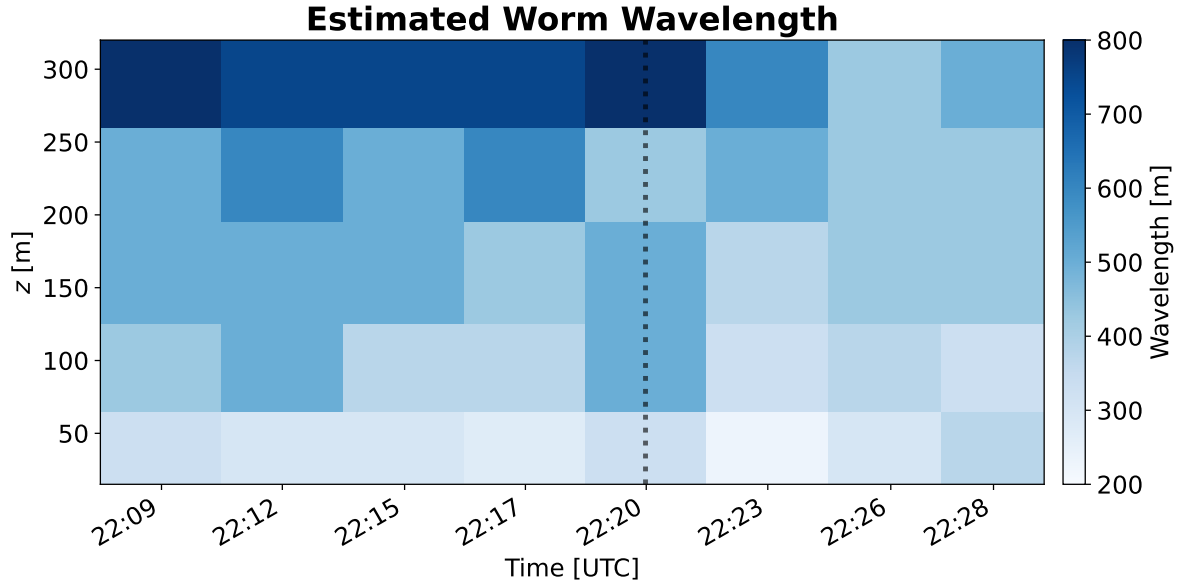


Figure 4.5: Time-height series of the wavelength of vorticity worms near OU-PRIME performed by subjectively counting worms that intersect the 3 km range black line from Fig. 4.4 over the lowest 5 sweeps. The vertical gray dotted line is as in Fig. 4.2.

While the analyses in Fig. 4.4 and 4.5 provide a time-height profile of wavelength of

the finest resolved vorticity worms in storm I1’s inflow, they do not provide insight on the nature of vertical vorticity near the mesocyclone itself. To do this, we employ a 2D multi-level Haar wavelet transform on the ζ_{inf} field (discussed in Sec. 3.3) to provide insight on the wavelength of vorticity structures in the near-field inflow and outflow regions of storm I1. It is important to understand that the power from the wavelet transform is proportional to the variance of the field it is calculated on, not just the magnitude. Indeed, partitioned time-height series of vertical vorticity variance from DDA show higher values in the outflow than inflow for nearly all vertical levels analyzed (Fig. 4.7). This would indicate that one could generally expect larger wavelet power in the outflow partitions herein than the inflow. Planar views of normalized wavelet power from the 0.25° sweep from the volume starting at 22:17 UTC are presented in Fig. 4.6. Power is normalized by the total power in a 4-km circle around the mesocyclone center (full detail in Sec. 3.4), so each grid cell represents a fraction of that total power on a log scale. Zonal/meridional power represents fluctuations of ζ_{inf} in that respective direction, i.e., a structure that is 500 m wide in the zonal direction but multiple kilometers long in the meridional direction (like inflow vorticity worms in this case) would have high zonal power and very little meridional power in the $400 < \lambda < 800$ m level (seen in Figs. 4.6b and e). In Fig. 4.6a and b, we generally see power > -25 dB in all regions of the storm, though it is noted that meridional power is lower in the inflow than the RFD region for $200 < \lambda < 400$ m wavelength level (Fig. 4.6d). This contrast is more pronounced at $400 < \lambda < 800$ m (Fig. 4.6b and e), as zonal power again appears high in all storm regions, but meridional power is much lower in the inflow than RFD. This quantitatively indicates that inflow vorticity structures are more elongated in the meridional direction, which visually aligns with Fig. 4.1g–i. In the RFD at wavelengths $400 < \lambda < 800$ m (Fig. 4.6b and e), larger power is seen in both directions south of the hook echo. At wavelengths $800 < \lambda < 1600$

m (Fig. 4.6c and f), meridional power is primarily located south and west of the hook echo, but zonal power is present in and north of the hook as well. This indicates that structures north of the mesocyclone in the RFD have more variability in the zonal direction than meridional for large wavelengths, but have similar variability at smaller wavelengths. Overall, the wavelet analysis indicates that vorticity structures in the RFD are primarily non-preferentially oriented (like vortex patches). In the inflow, zonal variability appears larger than meridional across all wavelengths, which is more representative of coherent streaks of ζ_{inf} oriented primarily north–south.

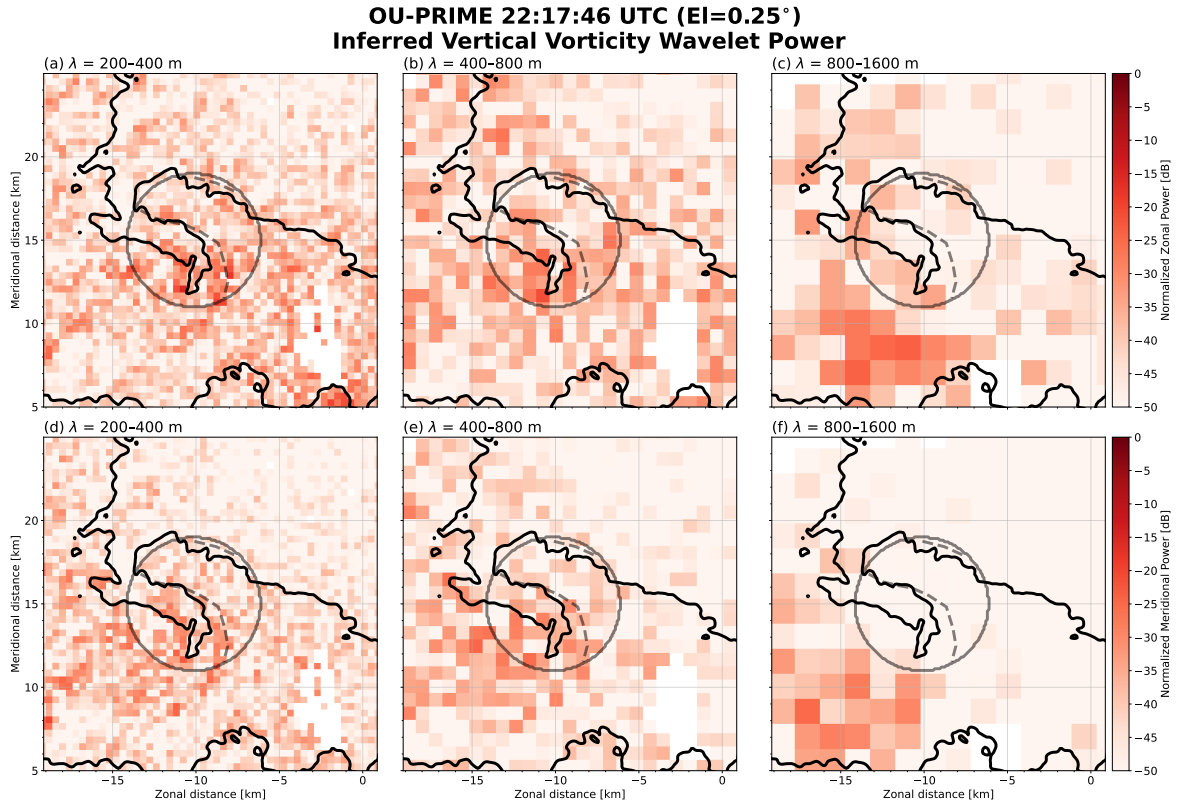


Figure 4.6: Zonal (a–c) and meridional (d–f) wavelet power spectrum of OU-PRIME ζ_{inf} (shaded) at 0.25° elevation from the volume starting at 22:17. The gray dashed line is the gust front and the solid gray line is the 4-km radius circle used for the partitioned time series analyses. Power is normalized by the total power across all levels within the 4 km radius circle used for the partitioned time series analysis and plotted on a log scale.

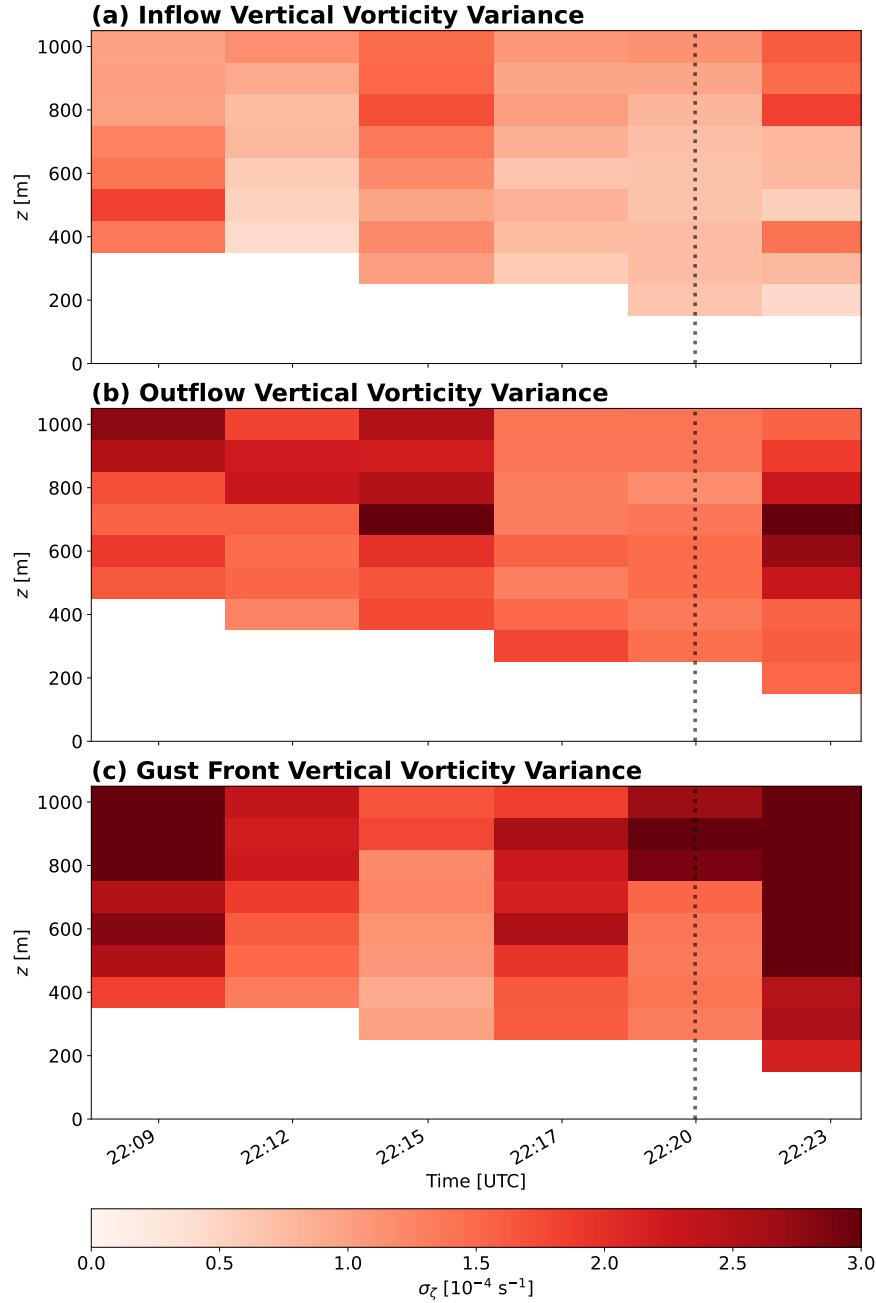


Figure 4.7: Time-height series of vertical vorticity variance (σ_ζ) in the inflow, outflow, and gust front partitions. Note a 4 km radius around the vortex position at each time is used. The vertical gray dotted line is as in Fig. 4.2.

We present wavelet power from the 2.0° sweep of the same volume in Fig. 4.8. For all wavelength levels, power is larger in the RFD than the far field inflow. Power

≈ -30 dB is seen in the near-field inflow for $200 < \lambda < 400$ m and $400 < \lambda < 800$ m, and power is generally larger in the zonal direction than meridional. The presence of wavelet power (and thus, variability in the ζ_{inf} field) close to the mesocyclone, but not the further field inflow, could be related to storm-scale processes or turbulence in the mesocyclone itself. This is because at 2.0° , the radar beam is $z > 500$ m AGL in the mesocyclone and $200 < z < 500$ m AGL in the inflow region, which is generally too high to observe vorticity worms. The larger power in the RFD shows that vorticity structures are present above the surface layer, in contrast to the inflow. An interesting feature is seen in Fig. 4.8c and f, where higher power is seen south of the gust front and west of the mesocyclone in both directions at wavelengths $800 < \lambda < 1600$ m. Though it is unclear what could cause higher power above the surface layer in this region west of the mesocyclone, we note that this region may not be entirely comprised of outflow air, as it is further separated from the RFGF.

Time series of summed wavelet power for the inflow partition are presented in Fig. 4.9 for the 0.25° , 1.0° , and 2.0° elevation angles. For all times, elevation angles, and directions, the $200 < \lambda < 400$ m and $400 < \lambda < 800$ m wavelength ranges have the largest power, strongly indicating that variance in the inflow ζ_{inf} is dominated by these spatial scales. At 0.25° , zonal power at these smaller scales ($\lambda < 800$ m) is around 0.04–0.05 before 22:15, before steadily decreasing to ≈ 0.01 at 22:23 (Fig. 4.9a). A similar pattern is seen in meridional power on this sweep, but power is slightly lower and decreases more sharply at 22:17. The power ratio (zonal / meridional) at 0.25° is close to one until 22:15, where it increases and remains positive through 22:23. This indicates that inflow ζ_{inf} initially had similar variability in both directions, but then variability decreased in the meridional direction, leading to the ratio exceeding one after 22:15. Similar trends are seen at 1.0° and 2.0° , where power is initially similar between the zonal and meridional directions, then zonal power significantly increases

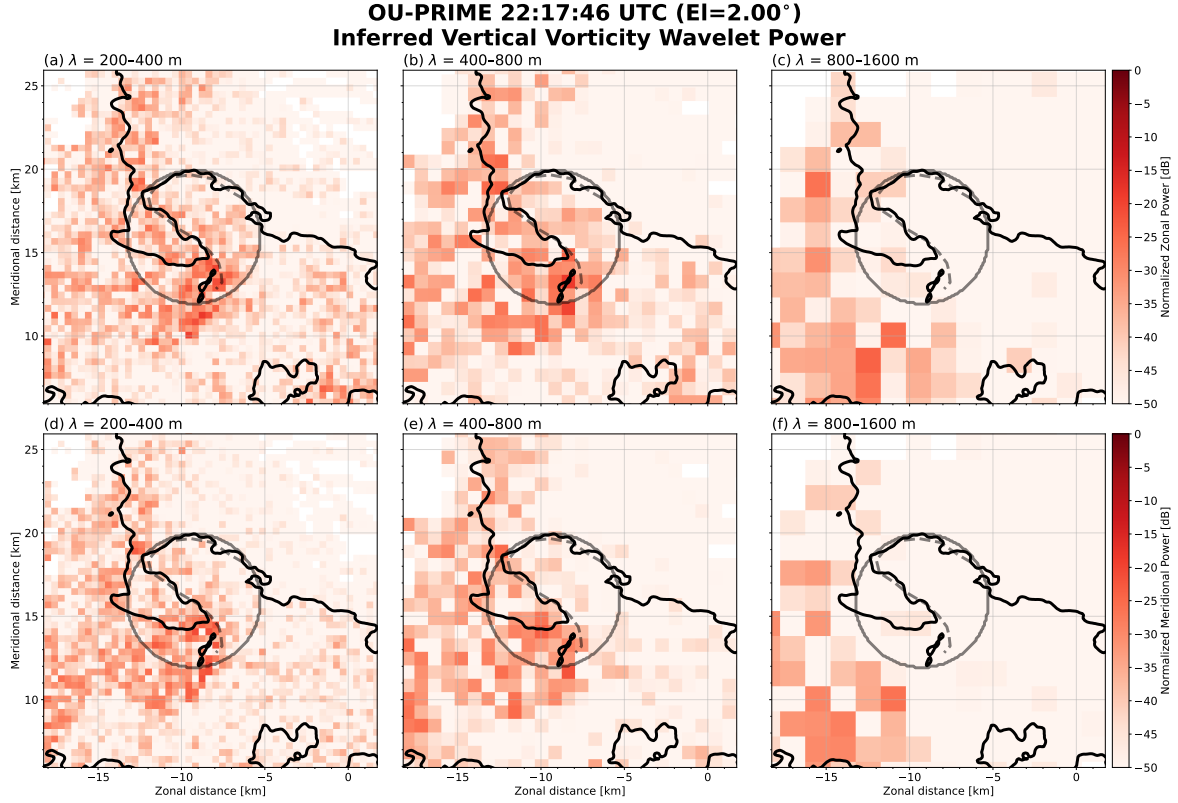


Figure 4.8: Same as Fig. 4.6 but for the 2.0° sweep.

to more than 0.04 at 22:20. Similar increases are seen in meridional power on these higher sweeps, though the magnitude is not as large as the zonal. Physically, this increase in the power ratio leading up to tornadogenesis indicates that ζ_{inf} structures became more uniform along the meridional direction, but maintained larger variability in the zonal direction. For longer, more coherent structures like vorticity worms, one would expect variability to be low along the worm as the value of ζ would remain near constant, but larger variability across the vorticity worm as ζ oscillates on the scale of hundreds of meters. Indeed, this is seen at the 0.25° elevation, as the power ratio becomes positive at 22:12 and increases to be over 2 (Fig. 4.9c), indicating variability in the zonal direction is double that of the meridional direction for all wavelengths analyzed. Additionally, the decrease in wavelet power in the inflow at 0.25° indicates

an overall decrease in variance of the ζ_{inf} field with time.

Inspecting OU-PRIME ζ_{inf} near the mesocyclone on the 0.25° sweeps supports this description. At 22:15, vorticity worms are seen in the inflow partition, primarily oriented southeast to northwest (Fig. 4.10a). At 22:20, coherent vorticity worms are seen along the RFGF and in the inflow, but are less defined within the inflow partition (Fig. 4.10c). While the ζ_{inf} field in the inflow appears less variable at this time (resulting in the lower summed wavelet power at this time), the structures seen do appear to be more north–south oriented, resulting in the positive power ratio seen in Fig. 4.9c. At 22:23, lower magnitude vorticity worms can be seen in the eastern part of the inflow partition (Fig. 4.10d). There is also a larger area of positive ζ_{inf} along the gust front, north of the hook echo, likely associated with the now ongoing tornado. Both of these structures are primarily north–south oriented, again explaining the positive power ratios at this time. Thus, these results indicate that surface layer vorticity worms in the near inflow congregated along the gust front and underneath the mesocyclone as tornadogenesis occurred, leading to an increase in their wavelength and an overall decrease in the ζ_{inf} field. It is also interesting that at 0.25° , zonal wavelet power decreases with time, but at 1.0° and 2.0° , power increases over time. This shows that while ζ_{inf} became less variable in the surface layer (0.25° elevation), it became more variable in the mixed layer (2.0° elevation). This indicates that structures in the surface layer and mixed layer are impacted differently as the mesocyclone and tornado vortex intensified.

Partitioned time series of summed wavelet power in the outflow are presented in Fig. 4.11 for the same sweeps as in Fig. 4.9. As in the inflow, power is largest in the $200 < \lambda < 400$ and $400 < \lambda < 800$ m wavelength levels for most times, though power is seen at larger wavelengths ($\lambda > 800$) m for the first few volumes analyzed. Power values are generally higher in the outflow than inflow, which, as mentioned previously, is likely due to there being more ζ_{inf} variance in the outflow (Fig. 4.7). For

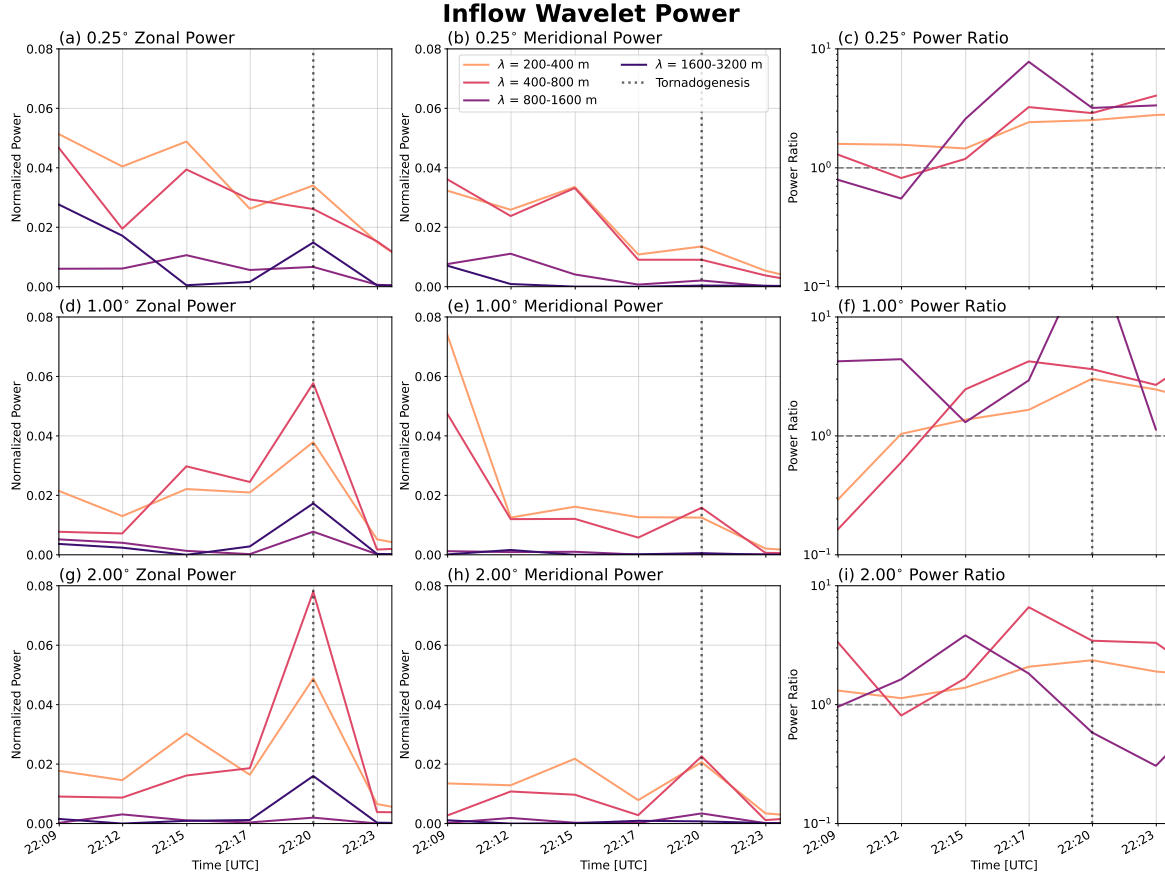


Figure 4.9: Time series of summed wavelet power in the inflow partition normalized by the total power in the circle across all levels. Wavelet power is split into the zonal (a,d,g) and meridional (b,e,h) components and the ratio of zonal to meridional power is presented in (c,f,i). Note that the final two OU-PRIME volumes are omitted as the supercell rapidly occluded after 22:23 UTC, leading to significant differences between the area of the inflow and outflow partitions. We also omit the $1600 < \lambda < 3200$ m wavelength level from the power ratios as it is very noisy. The mean radar beam height for the 0.25, 1.0, and 2.0° elevations is 105, 360, and 700 m AGL, respectively.

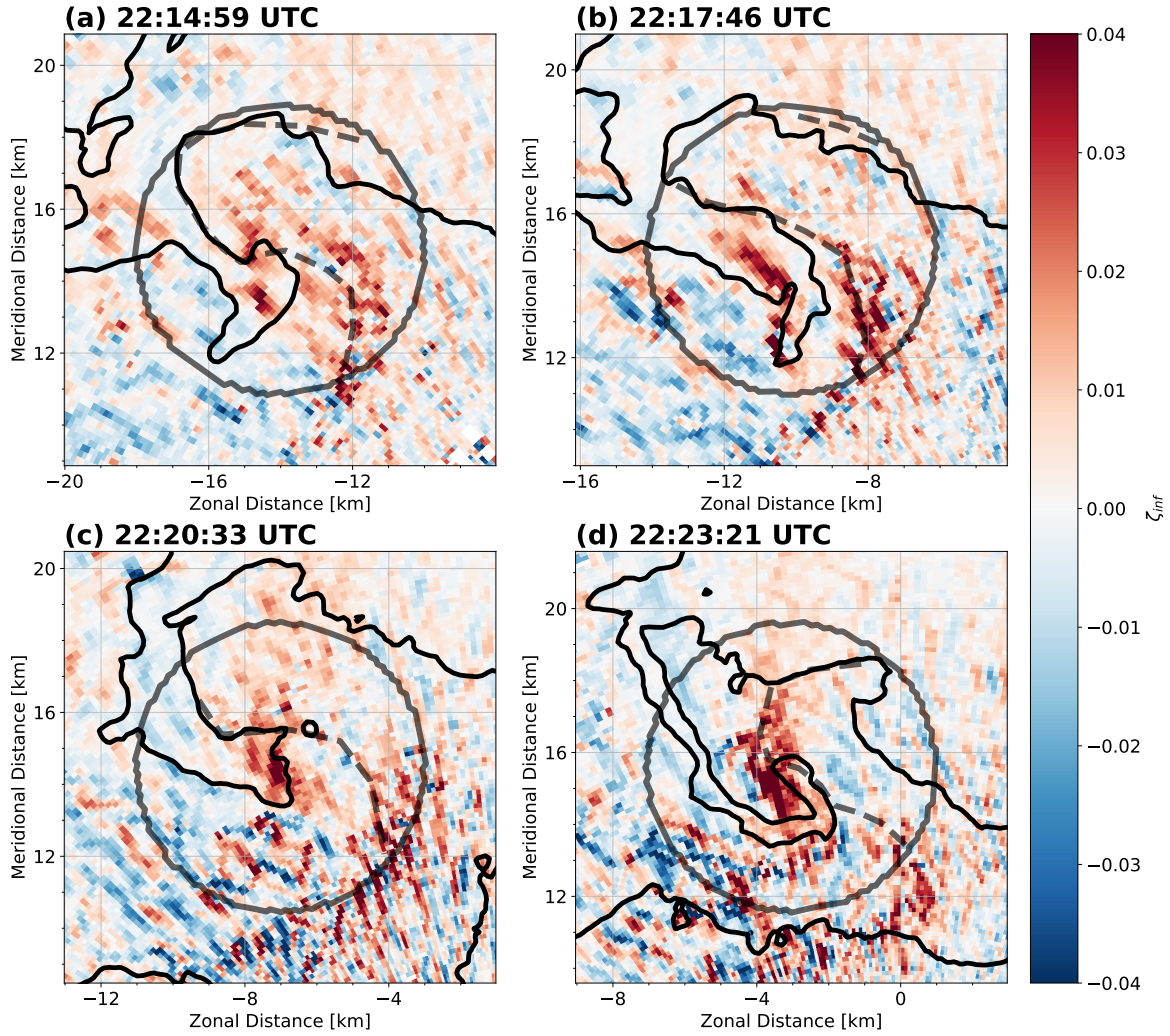


Figure 4.10: OU-PRIME ζ_{inf} at 0.25° elevation for four sweeps around the time of tornadogenesis. The solid black lines are reflectivity contoured at $Z = 20$ dBZ, the solid gray line is the 4-km radius circle, and the dashed gray line is the gust front used for the partitioned time series analysis.

all sweeps, power appears similar between all wavelength levels initially. This indicates the variability of the ζ_{inf} field is nearly uniformly dispersed across spatial scales. At 0.25° , zonal power for small wavelengths increases after 22:15 and approaches 0.20 by 22:23 while large wavelengths ($\lambda > 800$ m) remain between 0.02-0.05. In the meridional direction at 0.25° , power at wavelengths $1600 < \lambda < 3200$ m is very similar to the small wavelengths until 22:20, when it begins to decline. After 22:09, when power ratios at 0.25° are positive for most wavelengths, ratios approach one for two volumes before increasing back to above 1. The small wavelengths have a near constant, positive power ratio after 22:15. However, the large wavelengths appear much noisier. This could be due to there being less grid cells at large wavelength levels, lowering the number of samples in each partitioned region. The positive power ratio in the outflow partition may be due to the orientation of ζ_{inf} associated with the pre- and tornadic vortex itself, which is located within the hook and behind the gust front after 22:17 (Fig 4.10). During these times, while large magnitude ζ_{inf} covers a small spatial area, $\zeta_{inf} > 0.02$ extends primarily north and south from the tornado. This would yield an overall positive power ratio in this region. Additionally, large magnitude ζ_{inf} is found along the primary and secondary RFGF at 22:20 and 22:23, with many structures oriented north-south. At 1.0° and 2.0° , all wavelength levels have similar power magnitudes initially, then power decreases for the larger wavelengths and increases for smaller wavelengths. This would be indicative of outflow ζ_{inf} structures becoming smaller aloft as tornadogenesis occurred. Furthermore, for the 1.0° and 2.0° sweeps, the power ratios become greater than one between 22:15 and 22:17. This appears to be due to similar features that are seen at 0.25° (not shown).

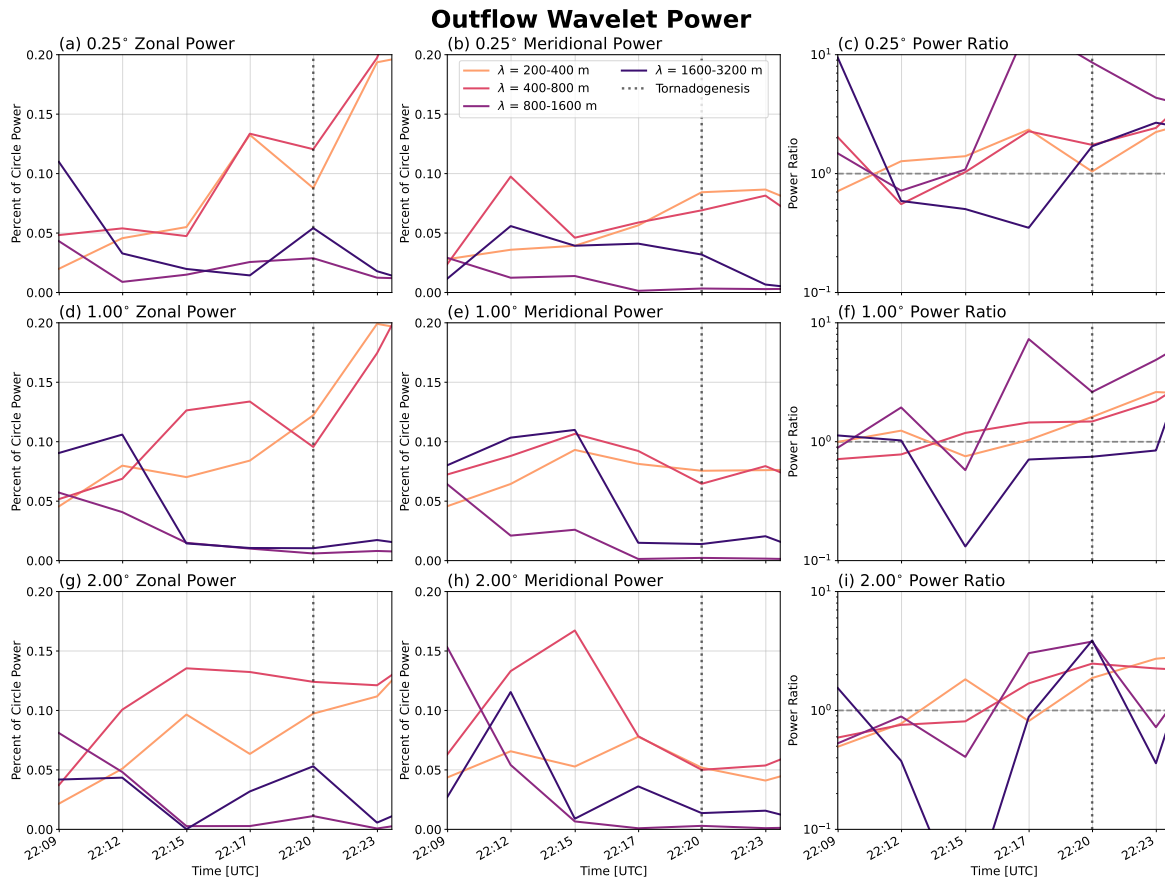


Figure 4.11: Same as Fig. 4.9 but for the outflow partition.

4.3 Available Vertical Vorticity in the Inflow and Outflow

While this dataset does not allow for analysis of which storm region provided the near-ground rotation for tornadogenesis, the high spatial resolution observations by OU-PRIME and DDA allow for a comparison of the vertical vorticity and circulation available in the inflow and outflow regions. Inspecting storm-relative winds and convergence from DDA with KTLX reveals a secondary RFGF at 22:15. This feature was noted in Palmer et al. (2011), but not until the 22:26 volume. The secondary RFGF is indicated by a second region of convergence behind the primary RFGF at 22:15

(Fig. 4.12a and e). In addition, when the second RFGF first becomes apparent, ζ_{inf} is generally of low magnitude and even broadly negative in this region of the RFD. At 22:17, larger sized and magnitude ζ_{inf} structures become apparent along and north of the secondary RFGF (Fig. 4.12c), and at 22:20, many structures with large magnitude ($|\zeta_{inf}| > 0.03 \text{ s}^{-1}$) are present (Fig. 4.12d and h). Furthermore, these structures on the secondary RFGF are larger and more patchy than the inflow vorticity worms and appear to be oriented NW–SE along the secondary RFGF. As the secondary RFGF formed, the enhanced convergence along it likely stretched vertical vorticity in the RFD, leading to the large ζ_{inf} magnitudes south and west of the hook echo at 22:20.

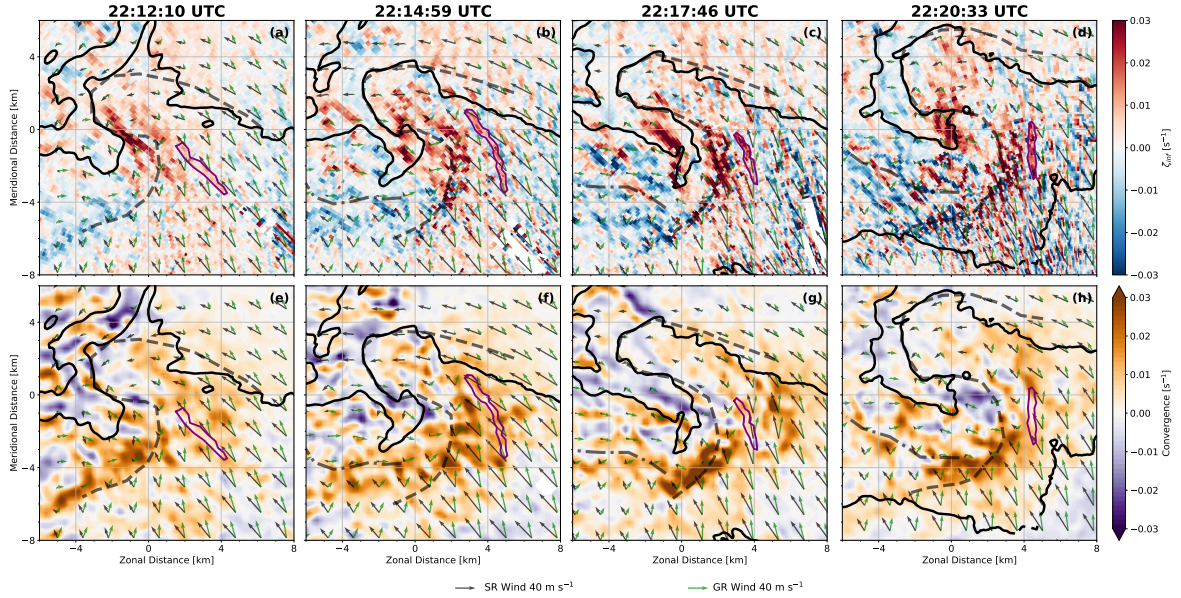


Figure 4.12: OU-PRIME $0.25^\circ \zeta_{inf}$ (a–d, shaded) and DDA convergence (e–h, shaded) for 4 volumes before tornadogenesis. The black and green vectors are storm-relative and ground-relative winds from DDA in m s^{-1} . Radial velocity is smoothed before calculating ζ_{inf} as in Fig. 4.1. The solid black lines are OU-PRIME reflectivity factor contoured at $Z = 20 \text{ dBZ}$. The dashed gray lines are the gust front as in Fig. 4.1 and the gray dashed and dotted lines are the secondary RFGF. All DDA variables are plotted at $z = 500 \text{ m AGL}$. The solid purple contours represent the subjectively identified vorticity worms analyzed in Tab. 4.1. Note that location labels are relative to the mesocyclone center.

During the 8-minute period plotted in Fig. 4.12, the magnitude of ζ_{inf} increases

significantly under the mesocyclone, along the primary RFGF, and in the near inflow. During this period, there is a large region of convergence $> 0.02 \text{ s}^{-1}$ along the RFGF and north and east of it, where the inflow decelerates and cyclonically curves into the mesocyclone. Following the storm-relative winds, vorticity worms would undergo stretching by the convergence along the RFGF, then be advected north in a storm-relative sense towards the hook. As this flow decelerates just east of the hook echo, vorticity worms continue to amplify and conglomerate, which is seen with the larger area of positive ζ_{inf} at 22:15 (Fig. 4.12b) near $x = 2, y = 1 \text{ km}$. Another interesting feature is the presence of two distinct areas of large ζ_{inf} starting at 22:15, one at the base of the hook, and a larger region along and on the warm side of the primary RFGF. The vortex in the hook appears to have originated from the vortex seen at 22:12, but was advected south by northerly storm-relative winds in this region where the inflow cyclonically curves (Fig. 4.12a). However, this is just speculation as our 2+ minute low-elevation updates prevent us from directly tracking this process.

Time	Area [m ²]	Average ζ_{inf} [s ⁻¹]	Circulation [m ² s ⁻¹]
22:12	$1.28 \cdot 10^6$	0.008	10691
22:15	$1.64 \cdot 10^6$	0.011	17761
22:17	$1.04 \cdot 10^6$	0.014	14120
22:20	$1.04 \cdot 10^6$	0.018	18329

Table 4.1: Area, horizontally averaged ζ_{inf} , and Doppler circulation of each vorticity worm identified in Fig. 4.12 at four different times.

It is also important to note the significant circulation values of individual vorticity worms. Similarly to Davies-Jones et al. (2020) and Gebauer et al. (2026), we calculate the Doppler circulation as

$$\Gamma = \int_A \langle \zeta_{inf} \rangle dA \quad (4.1)$$

where the angle brackets denote a spatial mean. We then calculate Doppler circula-

tion over each worm identified with purple lines in Fig. 4.12 at four different times (Tab. 4.1). For all four times, the identified Doppler circulation of the vorticity worms exceeds $10000 \text{ m}^2 \text{ s}^{-1}$, which is comparable to notable subtornadic circulations and some very narrow tornadoes. These values of circulation in each worm may provide a sufficient pressure drop at the surface to initiate stage 3 of the Fischer et al. (2024) tornadogenesis model and begin the cascade to tornadogenesis, especially if this circulation is converged to a smaller area. Furthermore, the large number of vorticity worms available in this supercell’s inflow provides additional circulation to that of the outflow alone. Unfortunately, due to the relatively coarse temporal resolution of this dataset (about 2 min), we cannot track the amplification of these individual vorticity worms over time.

In addition to the vertical vorticity present in the inflow of storm I1, immense horizontal vorticity was generated by friction near the ground (Schenkman et al., 2014; Roberts et al., 2016; Roberts and Xue, 2017). Using the Norman Oklahoma Mesonet Station (NRMN), which was positioned in the inflow of storm I1 during our entire analysis, and radial velocity measurements centered over the location of NRMN from the 5 lowest OU-PRIME sweeps, we can produce a vertical profile of horizontal wind speed before tornadogenesis to estimate horizontal vorticity. We calculate the horizontal wind as

$$V_h = \frac{V_r}{\cos\theta\cos\phi}, \quad (4.2)$$

where V_r is OU-PRIME radial velocity, θ is the elevation angle (0.25° , 1.0° , 2.0° , 3.0° and 4.0°), and ϕ is the angle between the OU-PRIME azimuth angle and 10-m AGL wind direction from the NRMN mesonet site (Gebauer et al., 2026). To make this finding more applicable to storm I1’s inflow region as a whole during the pretornadic period, we average 10 minutes of NRMN wind speed data and 36 OU-

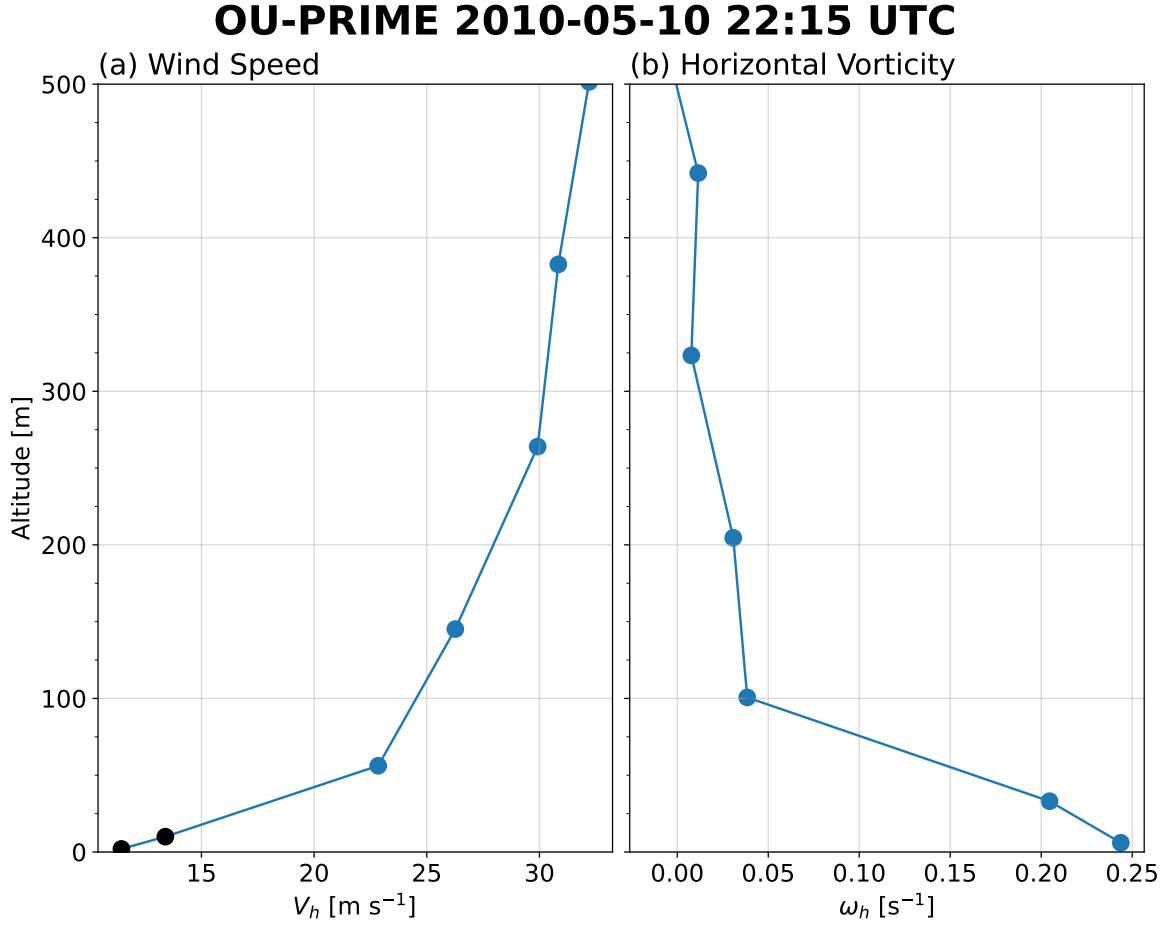


Figure 4.13: Vertical profiles of horizontal wind speed (a) from the NRMN Oklahoma Mesonet site (black dots) and OU-PRIME (blue dots) from the volume starting at 22:15 UTC and approximated horizontal vorticity (b). Note the wind speed estimates from NRMN are 10-min averages centered on 22:15 UTC while those from OU-PRIME are radial velocity averaged over 36 gates surrounding the NRMN site. This is done to represent a larger spatial region in the inflow.

PRIME radial velocity gates together from the volume starting at 22:15 to produce the vertical profiles. In the lowest 50 m AGL, horizontal vorticity exceeds 0.2 s^{-1} , then sharply decreases to below 0.05 s^{-1} at 100 m AGL (Fig. 4.13). This magnitude aligns well with the Markowski (2024) simulations and the observations shown in Gebauer et al. (2026). Due to the lower vertical resolution of our vertical profile in Fig. 4.13 than in Gebauer et al. (2026), which used sector velocity azimuth displays (VADs) instead of

multiple elevation scans, we cannot resolve the larger magnitudes of horizontal vorticity expected very near the surface (≤ 10 m AGL.). Nonetheless, the presence of $\omega_h \approx 0.25$ s^{-1} near the ground in the inflow of storm I1 poses immense tilting potential. The estimate of $\omega_h \approx 0.25$ is also applicable to the inflow of storm I1 as a whole, meaning this tilting potential is present over a larger region.

Furthermore, while we avoid directly inspecting horizontal vorticity in our DDA due to its dependence on vertical gradients, we can estimate the streamwiseness of frictionally generated horizontal vorticity in the inflow by comparing the angles of the storm-relative and ground-relative winds in DDA. One could infer that the horizontal vorticity in the lowest 500 m AGL would be approximately oriented 90° to the left of the ground-relative winds, meaning that pure streamwise vorticity would exist if the ground-relative winds are oriented 90° to the right of the storm-relative winds. During all times plotted in Fig. 4.12, the ground-relative winds (green vectors) are $30\text{--}90^\circ$ to the right of the storm-relative winds (black vectors), indicating that a significant portion of the frictionally generated horizontal vorticity near the ground is streamwise. In the portion of the inflow east and south of the RFGF, ground-relative winds exceed 40 m s^{-1} . This large magnitude would indicate that even if the streamwiseness of horizontal vorticity is only $\approx 30\%$, a large magnitude of streamwise vorticity would still exist. At 22:17 and 22:20 (Fig. 4.12c,g and d,h), the angle between the ground and storm-relative winds increases to be closer to 90° along the RFGF and in the broad region of convergence north and east of the hook echo. This highlights an abundance of streamwise vorticity in the near inflow of storm I1 during tornadogenesis in the lowest 500 m AGL resulting from friction, in addition to the vertical vorticity associated with vorticity worms in the surface layer.

While the 2-min time resolution of low-elevation sweeps in the OU-PRIME dataset prevents us from tracking individual vorticity worms, the DDA with KTLX can be

used to provide insight into how much near-ground circulation originated in the inflow and outflow during tornadogenesis. Using the partitioned time-series, we can compare the circulation of the inflow to the outflow summed over a 4-km radius circle centered on the mesocyclone at each time and elevation. The inflow partition is generally found to have larger or equal circulation values to the outflow in Fig. 4.14. Due to KTLX being further away from storm I1 and using a slightly higher elevation angle, we cannot characterize circulation below 200 m AGL until after tornadogenesis. However, it is still interesting that the low-level inflow had a similar or greater circulation than the outflow in this case. Greater circulation is seen in the gust front partition, which is indicative of the convergence and stretching of vorticity noted in Figs. 4.1 and 4.12.

While circulation content in the inflow and outflow partitions is similar during most of the tornadogenesis process, the storm-relative winds were seen to be much stronger in the inflow in Fig. 4.12. One could expect that with stronger storm-relative winds, given similar circulation, there would be more opportunity for vorticity contributing to the tornado vortex to be sourced from the inflow. To examine this, we calculate the flux of vertical vorticity by the horizontal winds into each circle partition as

$$F_{\zeta} = - \iint_A \left(u \frac{\partial \zeta}{\partial x} + v \frac{\partial \zeta}{\partial y} \right) + \zeta \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) dA, \quad (4.3)$$

where A is the area of the partition (inflow, outflow, or gust front). Partitioned time-height series of F_{ζ} reveal drastic differences between the inflow and outflow partitions, with entirely positive flux of vertical vorticity into the inflow partition and nearly all negative flux in the outflow. This sign difference between the inflow and outflow is most certainly due to the storm-relative winds, as they are near zero or easterly in the RFD in Fig. 4.12e–h, preventing vertical vorticity structures from interacting with the low-level mesocyclone. In contrast, the storm-relative winds in the inflow broadly

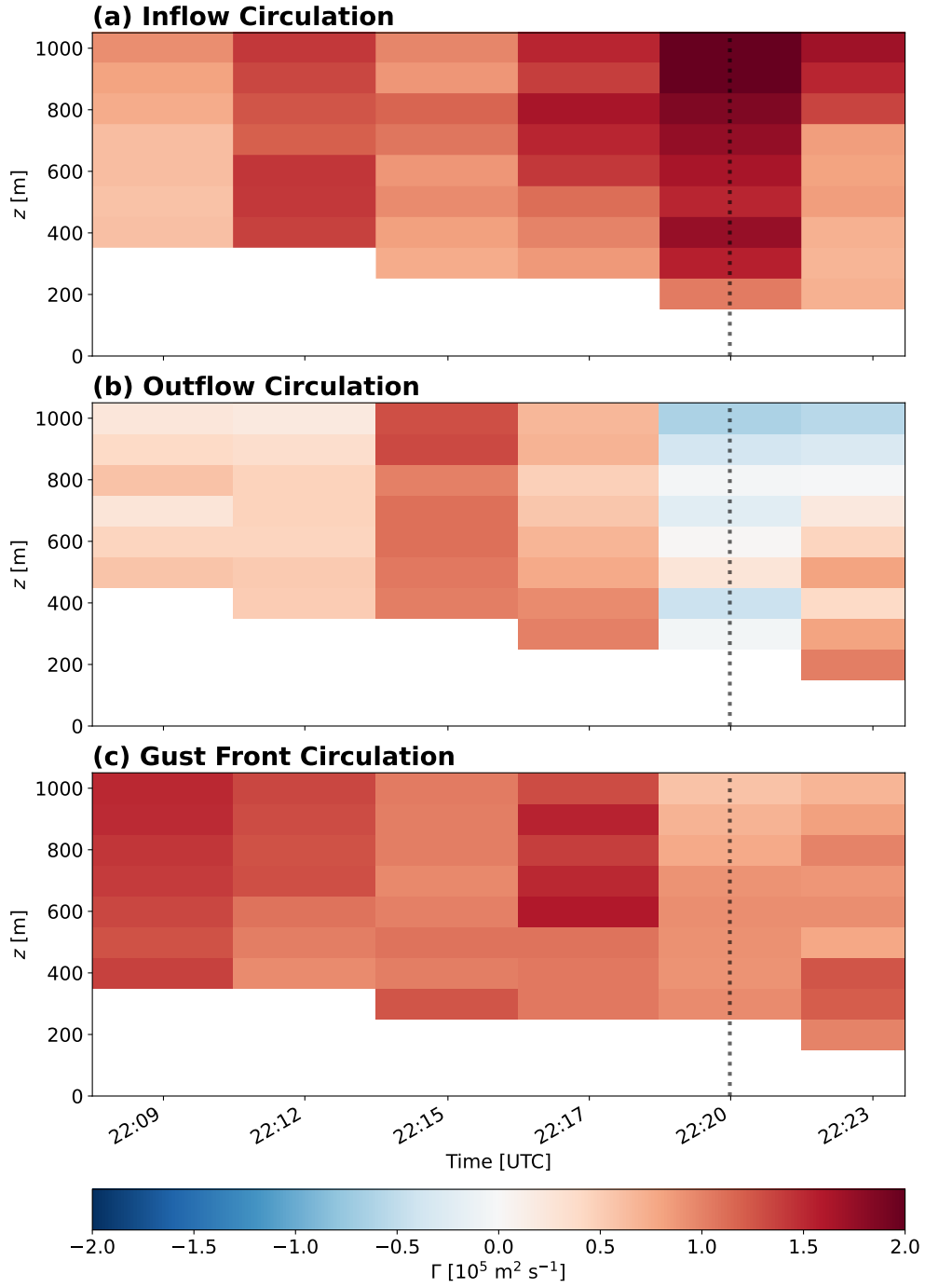


Figure 4.14: Same as Fig. 4.7, but for circulation.

exceed 50 m s^{-1} and are oriented such that vorticity along the RFGF is advected north and west towards the mesocyclone. F_ζ is generally of lower magnitude in the gust front

partition and includes both positive and negative values. This is likely because F_ζ only includes the advection and convergence of pre-existing vertical vorticity and not the tilting and stretching of vorticity.

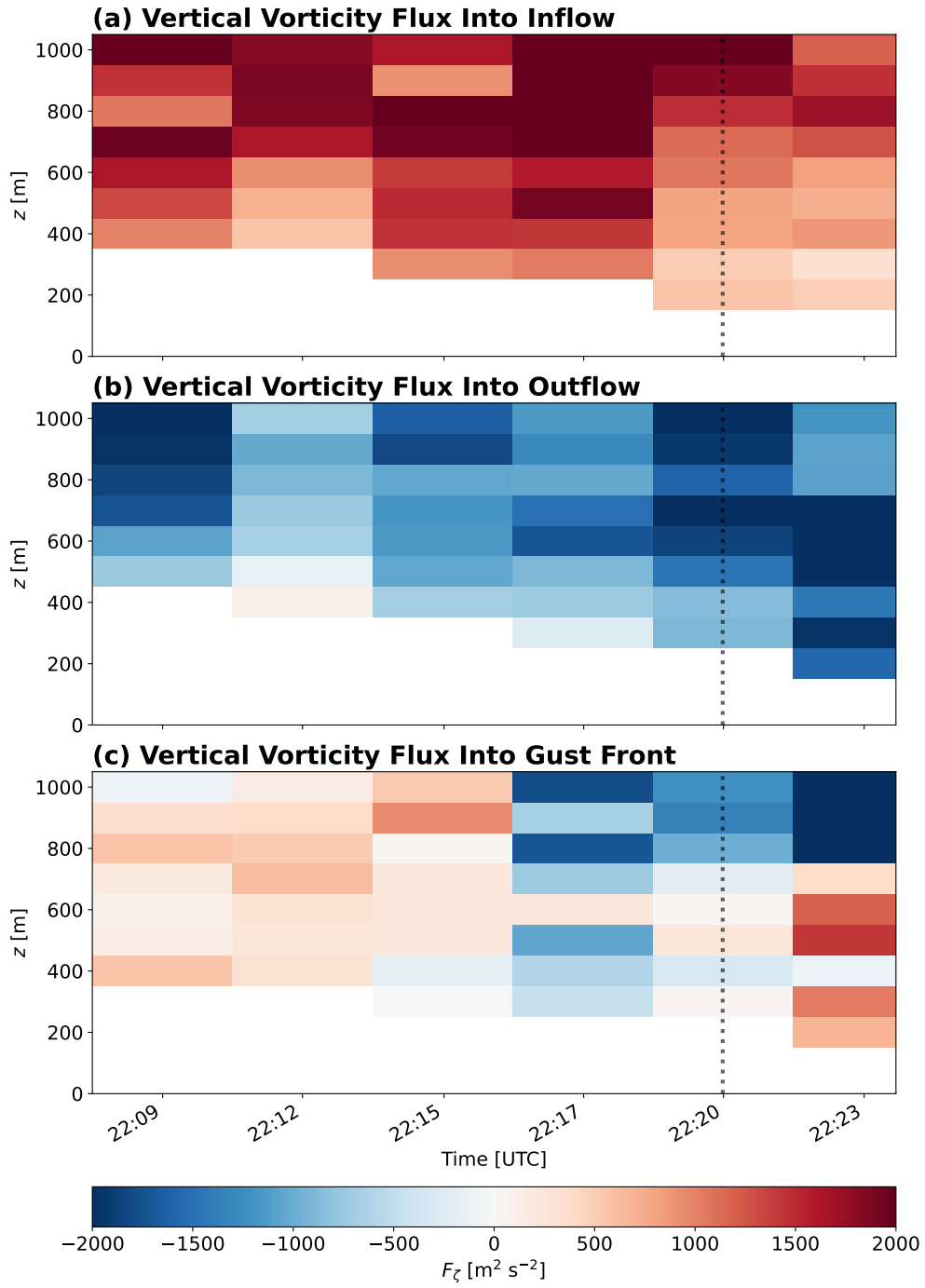


Figure 4.15: Same as Fig. 4.7, but for the flux of vertical vorticity into each partition by the horizontal wind from DDA.

Chapter 5

Discussion

5.1 Physical Interpretation of Tornadogenesis in Storm I1

While this dataset is insufficient to fully answer if vorticity worms facilitate tornadogenesis in real supercells, many interesting features are found in storm I1 that allow for speculation of the tornadogenesis process within it. The storm-relative winds in the inflow are generally much larger than in the RFD in Fig. 4.12 and are primarily southeasterly. As the primary RFGF advances east and south of the hook, this orientation of the inflow storm-relative winds seems to allow vorticity present in the inflow to advect northward to under the mesocyclone, rather than south and west of it. In contrast, while the westerly ground-relative winds were strong in the RFD ($20\text{--}30\text{ m s}^{-1}$), the storm-relative winds were close to zero, and even easterly at 22:15 and 22:20. Though we cannot directly measure this due to the 2+ minute temporal updates of 0.25° sweeps in this case, one could conclude that the weak storm-relative flow in the RFD may prevent vertical vorticity formed by storm-generated downdrafts from moving beneath the low-level mesocyclone during tornadogenesis in this supercell. Conversely, the extreme ($> 60\text{ m s}^{-1}$) storm-relative winds in the inflow, which were oriented in such a way to allow inflow vorticity to be advected into the mesocyclone, were better configured to allow vertical vorticity from environmental turbulence to pass beneath the low-level

mesocyclone.

This interpretation is further supported by the flux of vertical vorticity by the horizontal winds (F_ζ , Fig. 4.15) into the 4-km radius analysis circle centered on the mesocyclone. At all heights in the DDA with KTLX, F_ζ is positive for all times before tornadogenesis in the inflow. Conversely, F_ζ is negative for nearly all heights and times analyzed before tornadogenesis in the outflow. Given the above speculation regarding the weak storm relative winds in the RFD possibly preventing downdraft generated vertical vorticity near the surface from passing under the mesocyclone of storm I1, the finding of negative flux of vertical vorticity into the outflow partition is intuitive. This finding could highlight a sensitivity of tornado formation to the angle between the RFGF and the storm-relative winds in the inflow.

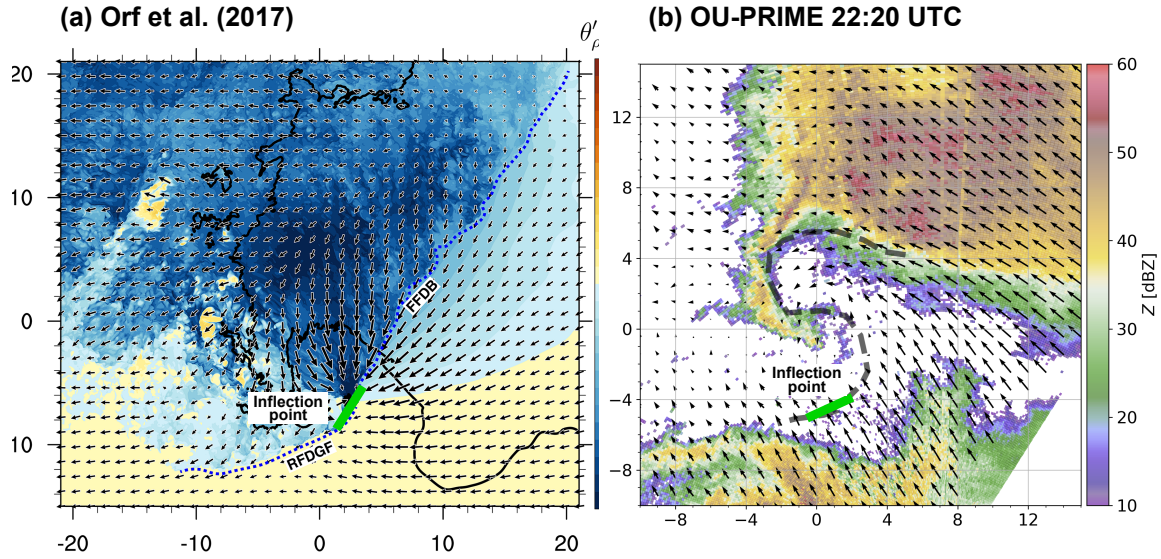


Figure 5.1: (a) Surface θ'_ρ (shaded), surface storm-relative winds (vectors), and 500 m AGL reflectivity contoured at $Z = 40$ dBZ (black line) with gust front boundaries denoted from the Orf et al. (2017) high resolution simulations of a violently tornadic supercell (their Fig. 5). (b) OU-PRIME 0.25° elevation reflectivity factor (shaded), 500 m AGL storm-relative winds from DDA (vectors), and the gust front used for the partitioned time series analysis (dashed gray line). The inflection points (green lines) denote where storm-relative inflow winds are perpendicular to the gust front.

Fig. 5.1 compares two low-level storm-relative flow patterns in different tornadic supercells. On the left is the tornadic supercell simulated by Orf et al. (2017), and on the right is storm I1 at 22:20 UTC observed by OU-PRIME and KTLX. In the Orf et al. (2017) supercell, much of the storm-relative winds are easterly and some even have a northerly component. The forward flank precipitation also has more of a southward extent than in storm I1, meaning much of the storm-relative inflow passes through precipitation before entering the mesocyclone. Following the storm-relative winds, one can identify an inflection point along the RFGF where the storm-relative inflow is perpendicular to the orientation of the RFGF, and thus, inflow winds to its north (in a storm-relative sense) can enter the mesocyclone, and flow to the south is deflected to the south and down the RFGF, west of the mesocyclone. In Orf et al. (2017), this inflection point is very near the mesocyclone due to the predominantly east northeasterly storm-relative inflow. In contrast, in storm I1 (Fig. 5.1b), the inflection point is nearly due south of the hook echo, leaving a larger area along the RFGF where storm-relative inflow can advect vorticity north into the mesocyclone. For storms oriented like in Orf et al. (2017), the angle between the storm-relative inflow and RFGF could prevent inflow vorticity sources from passing under the mesocyclone, meaning surface vertical vorticity for tornadogenesis must originate from the supercell's downdraft region. Conversely, the orientation and strength of the storm-relative inflow in storm I1 could better support inflow vorticity passing beneath the mesocyclone. Furthermore, the storm-relative winds north of the mesocyclone are very weak in storm I1 compared to Orf et al. (2017), where the northerly storm-relative winds are of similar strength to those in the inflow and likely better support interactions between vorticity generated in downdrafts and the mesocyclone. It is important to note this analysis is speculation based purely on the two snapshots presented in Fig. 5.1 and does not account for any temporal variation in the near-ground flow field around these supercells.

While the findings presented herein provide interesting observations of the near-ground vertical vorticity available in a tornadic supercell, it is important to note the shortcomings of this analysis and further work to be considered. First, it must be noted that this is only a study of one case in a highly volatile environment and the findings herein cannot be simply applied to all supercells. This is compounded by multiple uncommon features found in storm I1, such as the extreme ($> 40 \text{ m s}^{-1}$) inflow winds observed, which likely influenced the structure of the surface-layer in the inflow. In addition, the 2+ minute OU-PRIME volume update time (and ≈ 5 minutes for KTLX) prevents the tracking of individual vorticity structures near the ground during tornadogenesis in storm I1. These coarse volume update times likely also prevent accurate retrievals of variables that rely on vertical gradients within our DDA, preventing the use of analysis techniques like material circuits, backwards trajectories, and analyzing vertical vorticity tendency terms (Markowski et al., 2008, 2012; Kosiba et al., 2013). Capturing the evolution of all near-surface vorticity structures in a supercell during tornadogenesis, and ultimately answering if vorticity worms facilitate tornadogenesis in real supercells, would require rapid temporal ($\leq 30 \text{ s}$) updates at low altitude ($< 100 \text{ m AGL}$) in both the clear-air inflow and precipitation laden outflow regions Bluestein et al. (2010); Isom et al. (2013); Pazmany et al. (2013).

5.2 Environmental Impacts on Sources of Near-Ground Rotation for Tornadogenesis

Though the findings of Markowski (2024), Gebauer et al. (2026), and those herein are compelling and at least support that the inflow can provide similar amounts of near-ground vertical vorticity to the outflow in supercells, a plethora of observational research exists indicating the baroclinic vorticity generation within storm outflow re-

gions is a major source of near-ground rotation for tornado formation (Markowski et al., 2008, 2012; Kosiba and Wurman, 2013). As mentioned in Markowski (2024), it seems that supercells produce tornadoes in a wide variety of mechanisms, and improved numerical and observational techniques are just now discovering this diversity. As discussed herein, the source of near-ground vertical vorticity for tornadogenesis could be sensitive to the orientation between storm-scale boundaries and the storm-relative winds close to the surface. In storm I1 (Chap. 4), the southerly component of the inflow storm-relative winds and curved orientation of the RFGF may have allowed vertical vorticity in the inflow to advect into the low-level mesocyclone. In contrast, many simulated supercells, like Orf et al. (2017), have predominantly easterly storm-relative winds near the ground in the inflow, seemingly preventing vertical vorticity in the inflow from interacting with the mesocyclone. This raises the question: Which environmental conditions control the orientation of storm-scale boundaries and storm-relative flow near the ground around the low-level mesocyclone? RFDs, which lead to the formation of RFGFs, can form as the result of negative buoyancy (evaporative cooling, melting hail, and precipitation loading), vertical gradients of vertical vorticity, and vertical pressure gradients (Markowski, 2002). Many studies highlight the production of negative buoyancy at mid-levels as a primary instigator of RFDs (Browning and Ludlam, 1962; Browning and Donaldson, 1963; Lemon, 1977; Barnes, 1978; Klemp et al., 1981) and that mid-level storm-relative flow can influence the rate of evaporative cooling (Barnes, 1964). Other studies indicate that an occlusion downdraft, which is a smaller scale downdraft close to the surface behind the RFGF, forms dynamically as the result of intensifying low-level circulations, which drive dynamic perturbation pressure gradients near the ground (Klemp and Rotunno, 1983; Wicker and Wilhelmson, 1995). Given these findings, one could conclude conditions supporting large rates of evaporative cooling in the mid-levels and strong low-level circulations would support

rapidly advancing RFGFs.

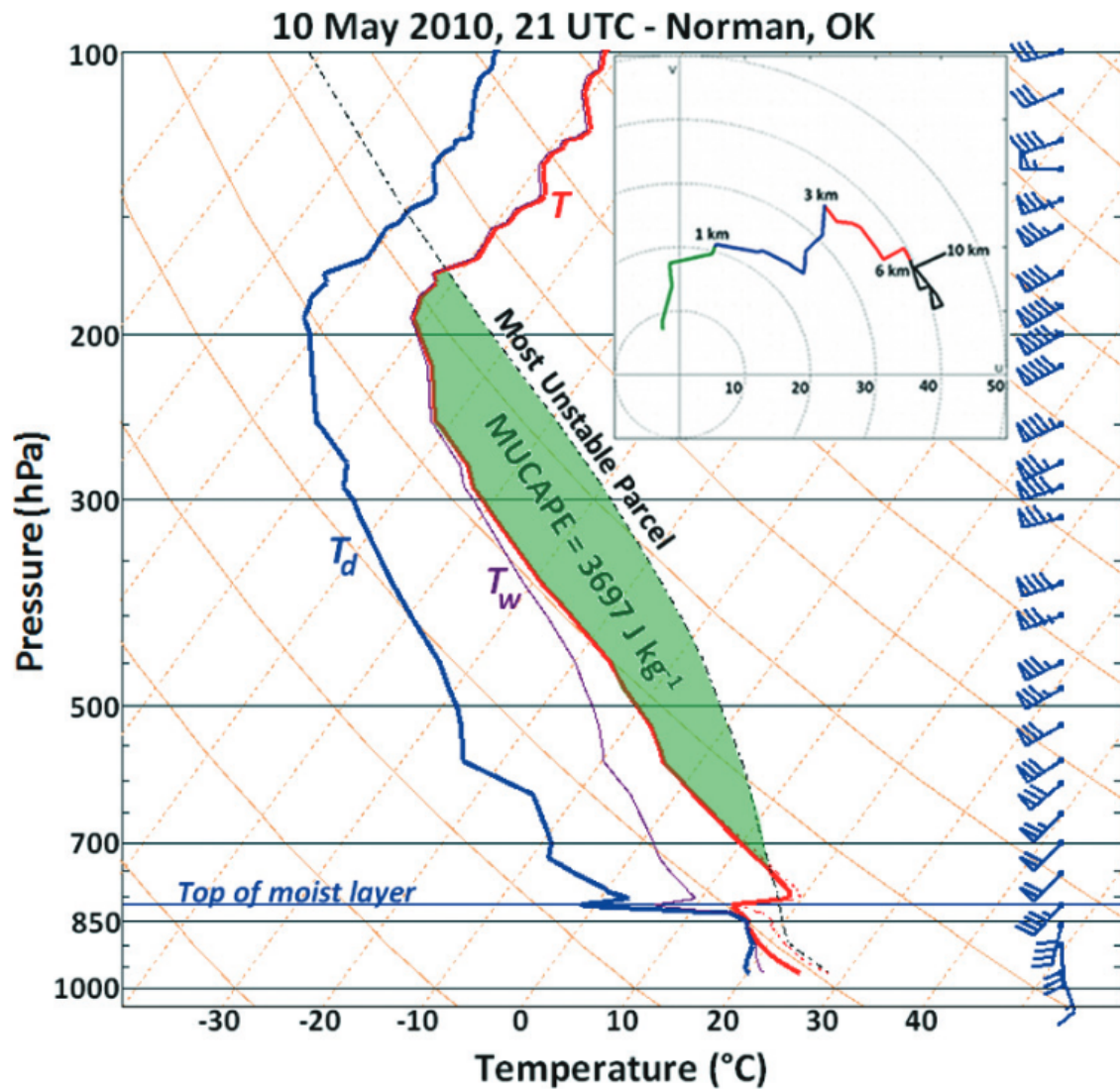


Figure 5.2: 10 May 2010 21 UTC sounding released by NWS Norman from Fig. 2 in Palmer et al. (2011).

Environments consisting of higher boundary layer relative humidity are associated with lower cloud bases and more buoyant cold pools than environments with lower boundary layer relative humidity and higher clouds bases (Markowski, 2002). The environment in central Oklahoma on 10 May 2010 featured a very moist boundary layer, with high relative humidity present from the surface to above the 850 hPa level

(Fig. 5.2). Additionally, storm motions were near 50 knots on 10 May 2010 (Palmer et al., 2011). This combination of high boundary layer relative humidity and fast storm motions may have limited the advancement of the RFGF in storm I1, which seemingly did not occlude until after genesis of the Moore-Choctaw EF4 tornado (Fig. 4.3). This likely indicates that buoyant generation of the RFD was limited and dynamic forcing dominated the formation of the occlusion downdraft as the low-level circulation intensified. Further research is needed to determine which environmental conditions would support facilitation of tornadogenesis by vertical vorticity in the inflow or outflow.

5.3 Other Observations of Vorticity Worms

While storm I1 on 10 May 2010 is an excellent case to characterize near-ground vorticity in all regions of a tornadic supercell, it is important to inspect other supercells to ensure the findings herein are not unique to storm I1. The formation of a tornado near Littlefield, TX on 25 April 2025 was observed by the National Severe Storms Laboratory’s (NSSL’s) mobile Doppler Wind Lidar (DWL), which was apart of the the Low-level Internal Flows in Tornadoes (LIFT) experiment. The Littlefield, TX tornado was associated with a supercell that interacted with an outflow boundary (OFB, denoted by the lines of low reflectivity in the supercell’s inflow in Fig. 5.3), which may have played a role in the tornadogenesis process.

NSSL performed 60° sector scans with their DWL at 0.5° elevation every ≈ 30 seconds, providing better time resolution than the 10 May 2010 OU-PRIME dataset. As shown in Fig. 5.4, the DWL observed the RFGF, FFCB, and OFB during the formation of the Littlefield, TX tornado. Inflow vorticity worms were observed ahead (east) of the RFGF and appear to interact with the RFGF and OFB. At 20:51:14 UTC, a larger region of enhanced ζ_{inf} is observed north of the OFB (Fig. 5.4d) and at the

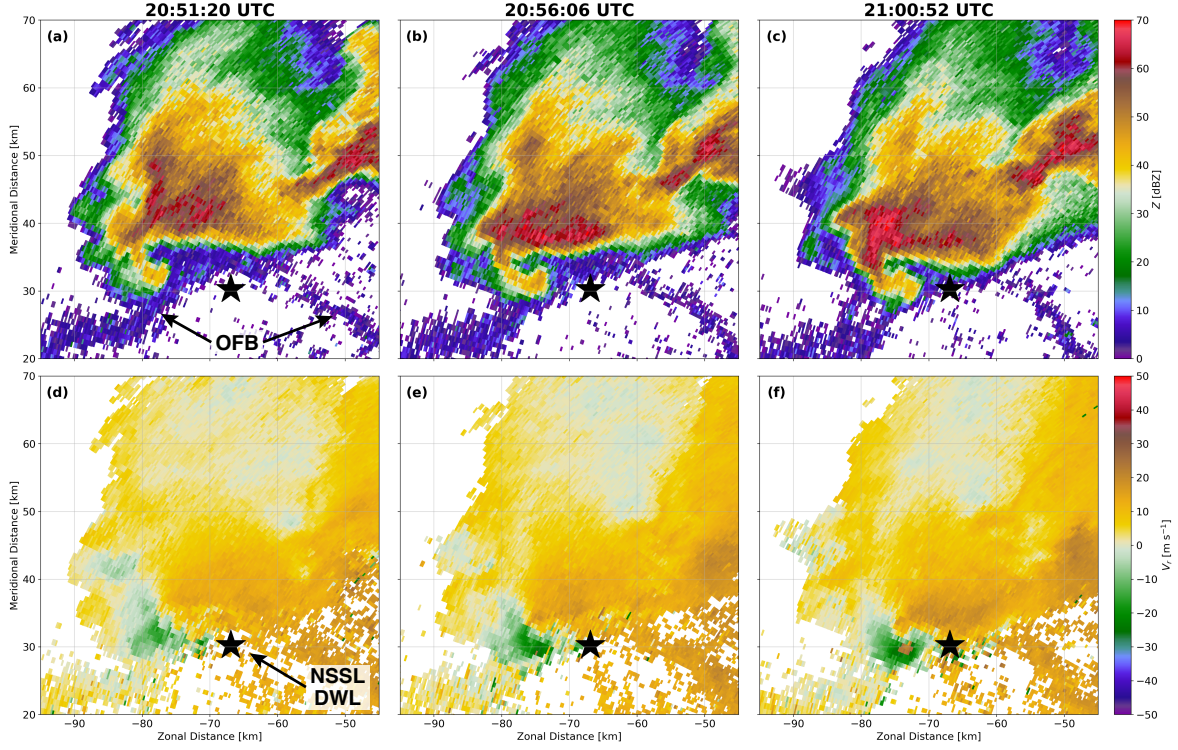


Figure 5.3: KLBB PPIs of reflectivity factor (a–c) and radial velocity (d–f) of the supercell and tornado observed by LIFT on 25 April 2025. The black star is the NSSL DWL deployment site.

same zonal position as a vorticity worm, which extends south-southeast of the OFB. Enhanced ζ_{inf} is also seen near the edge of the lidar range near $x = 7$, $y = 3$ km. This appears to be either just east of or underneath the mesocyclone at this time and is likely associated with a pre- or weakly-tornadic ground circulation. At 20:55:50 UTC, the hook echo from KLBB is now within the observable lidar range, and a weak tornado is now visible in the DWL radial velocities (Fig. 5.4b). A large region (≈ 2 km wide) of $\zeta_{inf} > 0.03$ is present associated with the tornado, and patchy ζ_{inf} structures extend east of the tornado along the RFGF and OFB. At 21:01:08 UTC, the tornado appears stronger with $\Delta V_{max} \approx 50$ m s⁻¹. The mesocyclone appears more occluded now, with the RFGF extending north of the tornado, then arcing east and then south. A secondary RFGF becomes apparent at this time, denoted by a small region of near zero

or slightly positive radial velocities west of a primary RFD surge. Large ζ_{inf} structures, similar to the vortex patches discussed in Fischer et al. (2024), are apparent along this second RFGF. Furthermore, these structures appear similar to those in Fig. 4.10 on 10 May 2010, with larger wavelengths and more random orientation.

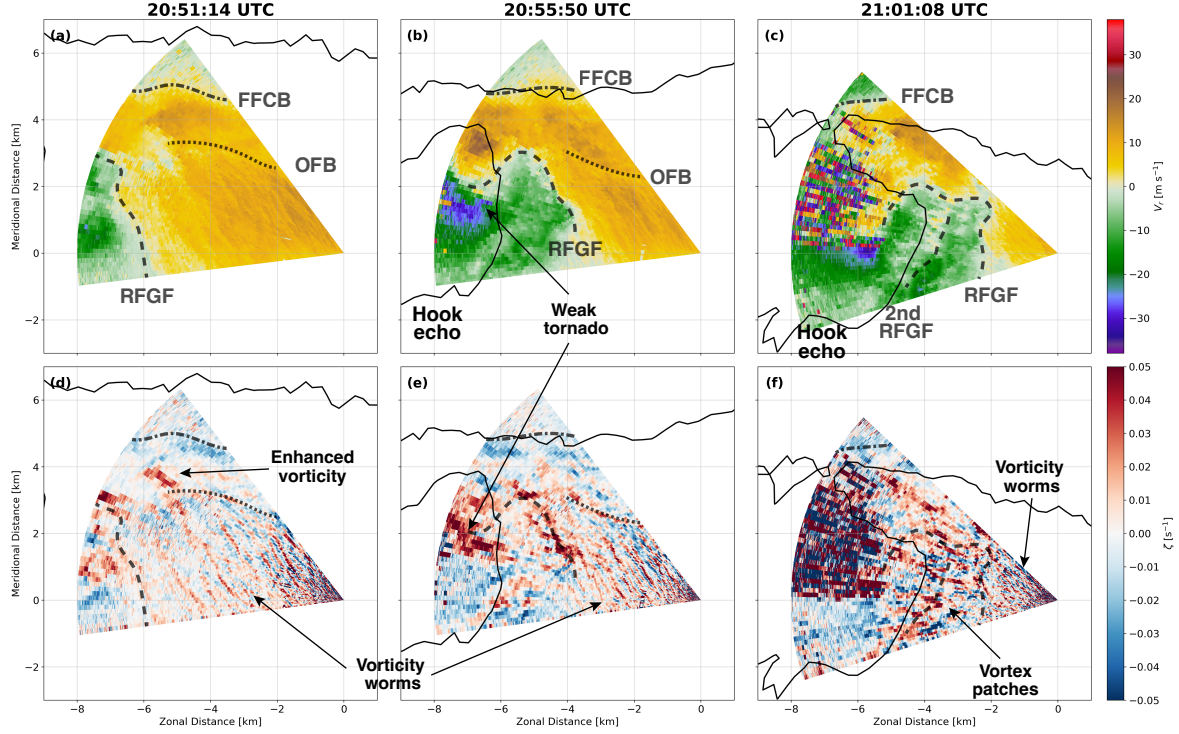


Figure 5.4: NSSL DWL 0.5° elevation PPIs of radial velocity (a–c) and inferred vertical vorticity (d–f). Solid black lines are KLBB reflectivity factor contoured at $Z = 20$ dBZ from the volumes in Fig. 5.3. The times listed are the middle time of the NSSL DWL PPIs plotted.

Three DWL PPIs from before tornadogenesis are presented in Fig. 5.5a–f. At 20:48:27 UTC, the low-level mesocyclone (LLMC) and RFGF are west of the observable lidar range, but the FFCB and OFB are visible. The enhanced vorticity structure along the OFB observed in Fig. 5.4d is first seen in the 20:48:27 UTC sweep, appearing to enter the DWL PPI from the east. This structure is located ≈ 1 km further east at 20:50:04 UTC (Fig. 5.5e), indicating a westward translation of $\approx 11 \text{ m s}^{-1}$. Furthermore, a vorticity worm appears to extend southeast from the enhanced ζ_{inf} north of

the OFB. It is uncertain if the enhanced vorticity structure originated from the inflow and amplified along the OFB, or if it originated from the storm itself. Regardless, it is interesting that ζ_{inf} associated with inflow vorticity worms appears attached to enhanced ζ_{inf} north of the OFB. If the enhanced ζ_{inf} north of the OFB did not originate from an inflow vorticity worm, one could speculate that perhaps the rotation within this structure generated a pressure drop, which then pulled the inflow vorticity worm towards it. Nonetheless, this enhanced vorticity structure increased in magnitude and size over the ≈ 3 minutes plotted in Fig. 5.5a–f as it approached the LLMC. Additionally, enhanced ζ_{inf} is seen along and behind the RFGF as it enters the observable lidar range at 20:51:55 UTC (Fig. 5.5f) and the weak/pre-tornadic circulation is visible at the edge of the lidar’s range.

Between 20:53:04 and 20:53:32, the enhanced ζ_{inf} structure originally found north of the OFB merges with the ground circulation near the intersection of the RFGF and OFB (Fig. 5.5j and k). A streak of positive ζ_{inf} extends east from the primary ground circulation at 20:53:32 and appears to have merged with the primary circulation by 20:54:13. Furthermore, over the ≈ 70 second period plotted in Fig. 5.5g–l, it appears that vorticity worms conglomerate near the intersection between of the RFGF and OFB. Given the mostly southeasterly ground-relative winds in the inflow (which can be inferred from the orientation of the vorticity worms), it appears that the vorticity in the inflow is translating along the RFGF and OFB towards the LLMC, where it converges.

As the ground circulation intensifies, vortices along the RFGF and within the RFD merge into the surface vortex (20:55:23–20:56:32, Fig 5.5p and q). As this convergence occurs, the radius of maximum winds (RMW) expands, and inbound radial velocities approach the DWL’s Nyquist velocity of 38 m s^{-1} . Around this time, a more clearly visual ground circulation with lofted dust and a funnel cloud become visible in a video

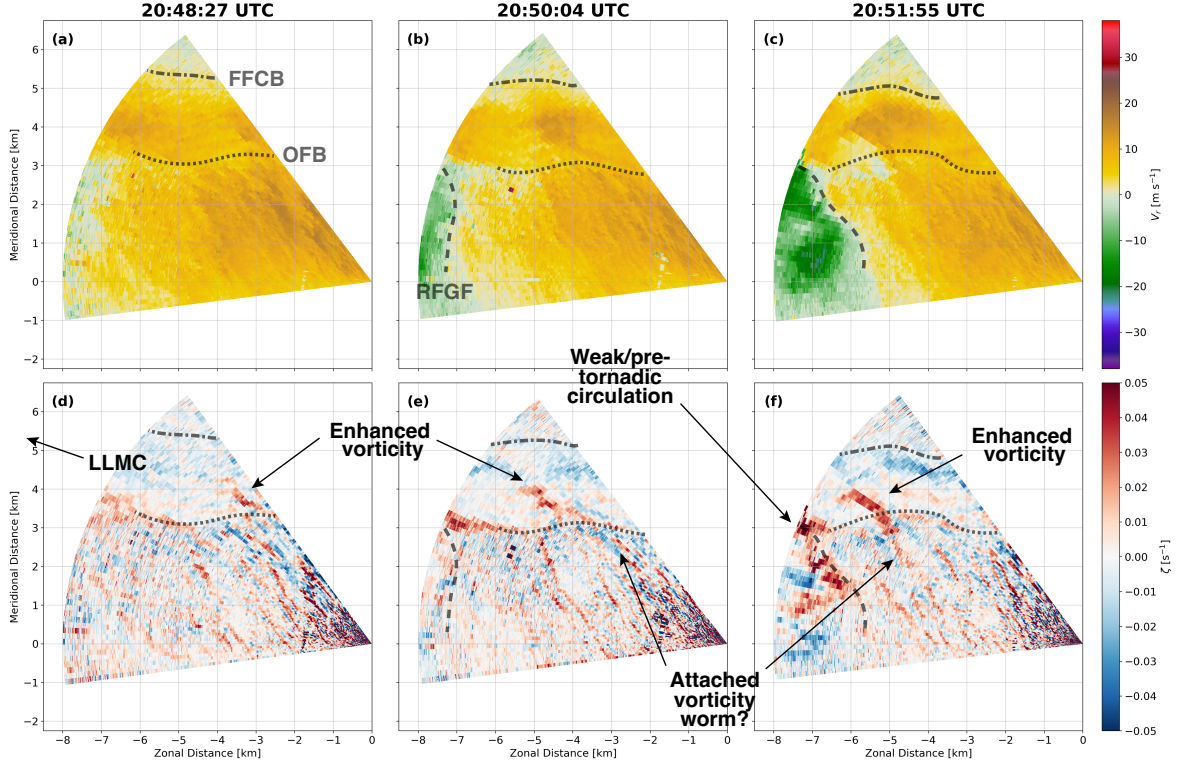


Figure 5.5: NSSL DWL PPIs of radial velocity (top row) and inferred vertical vorticity (bottom row). The times listed are the middle time of the NSSL DWL PPIs plotted. RFGFs are denoted with gray dashed lines, the OFB is denoted by gray dotted lines, and the FFCB is denoted by gray dashed and dotted lines.

taken from the NSSL DWL (not shown). The second RFGF becomes apparent during this intensification period (Fig. 5.5m-r), which is potentially evidence of an occlusion downdraft forming in response to the intensifying ground circulation. The region between the primary and secondary RFGF appears turbulent, with randomly oriented vortex patches present. Behind the secondary RFGF, the flow appears more laminar, with low magnitude ζ_{inf} present. East of the ongoing tornado, a second circulation forms along the RFGF. This circulation appears elongated, spanning ≈ 4 km from northwest of the OFB/RFGF intersection southeast into the inflow. As the LLMC occludes and the RFGF surges east, this secondary circulation appears to rotate around the tornado, advecting to the west into the cooler outflow air (not shown).

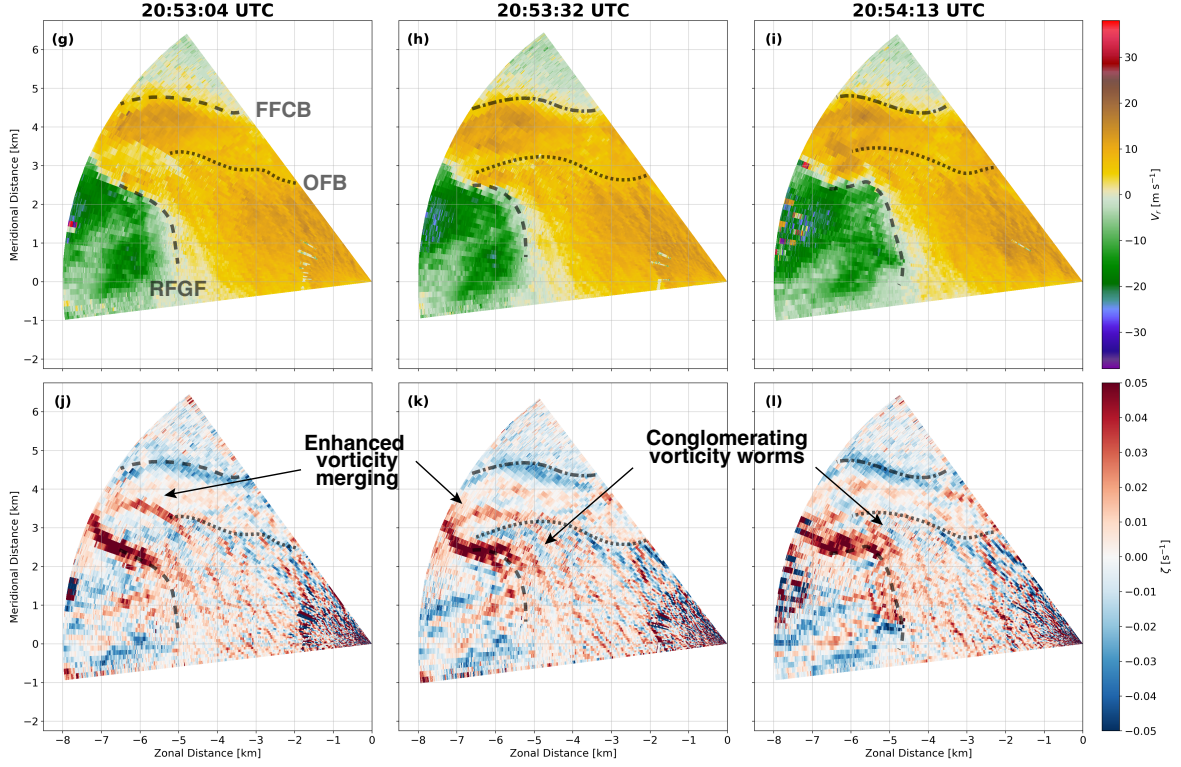


Figure 5.5: *(Continued)*

The observations by the NSSL’s DWL of the genesis of the Littlefield, TX tornado on 25 April 2025 provide an opportunity to analyze the amplification of vorticity worms near the ground at fine spatiotemporal resolution during the tornadogenesis process. Further analysis of this case and application of the wavelet transform, as performed herein on the 10 May 2010 case, would provide a quantitative measurement of the spatial scale of near-ground vertical vorticity in this case. Additionally, the ≈ 30 second temporal updates of the DWL observations allow for more direct examination of the influence of vorticity worms on the genesis of this tornado.

In addition to the OU-PRIME and NSSL DWL observations presented herein, many observations of vorticity worms in supercell inflow are present in existing literature. Bluestein et al. (2014) presented observations of the PBL near supercells obtained by collocated mobile Doppler radar and the Truck-Mounted Wind Observing Lidar Facility

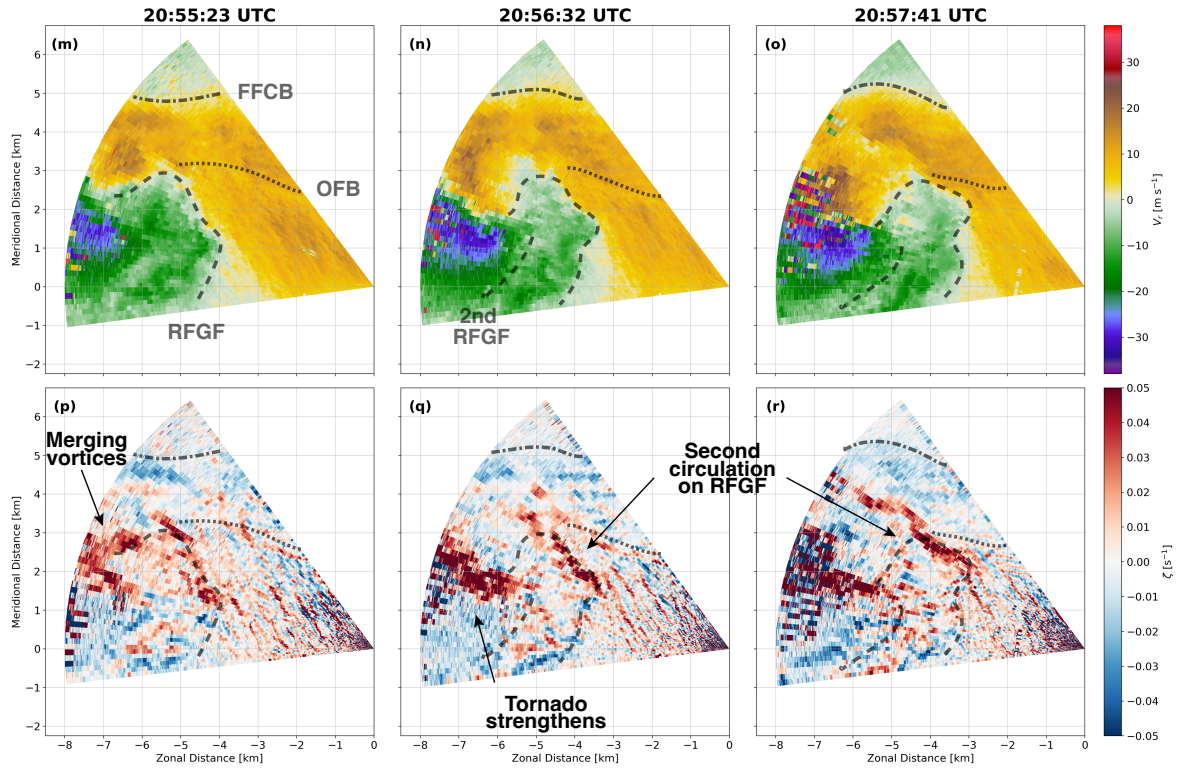


Figure 5.5: (*Continued*)

(TWOLF) during the Second Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX2). Their TWOLF PPIs of radial velocity and SNR from the inflow of a supercell on 19 May 2010 are presented in Fig. 5.6. High- and low-speed streaks are seen in the radial velocity field, and different spatial scales are seen with increasing range. The 4.0° elevation angle leads to a beam height of ≈ 500 m AGL at the gust front, which is located around a range of around 7 km in Fig. 5.6 and is denoted by the region of large SNR. The lidar beam appears to attenuate at the gust front, likely due to the presence of low clouds, and no velocity data are retrieved behind it. Within 3 km of the lidar, smaller low-speed streaks, on the order of hundreds of meters, are visible. At ranges greater than 3 km, the low-speed streaks appear wider and more similar to HCRs oriented southeast to northwest. At 4.0° elevation, the lidar beam reaches 200 m AGL around a range of 2.5 km. While the ASL depth in this

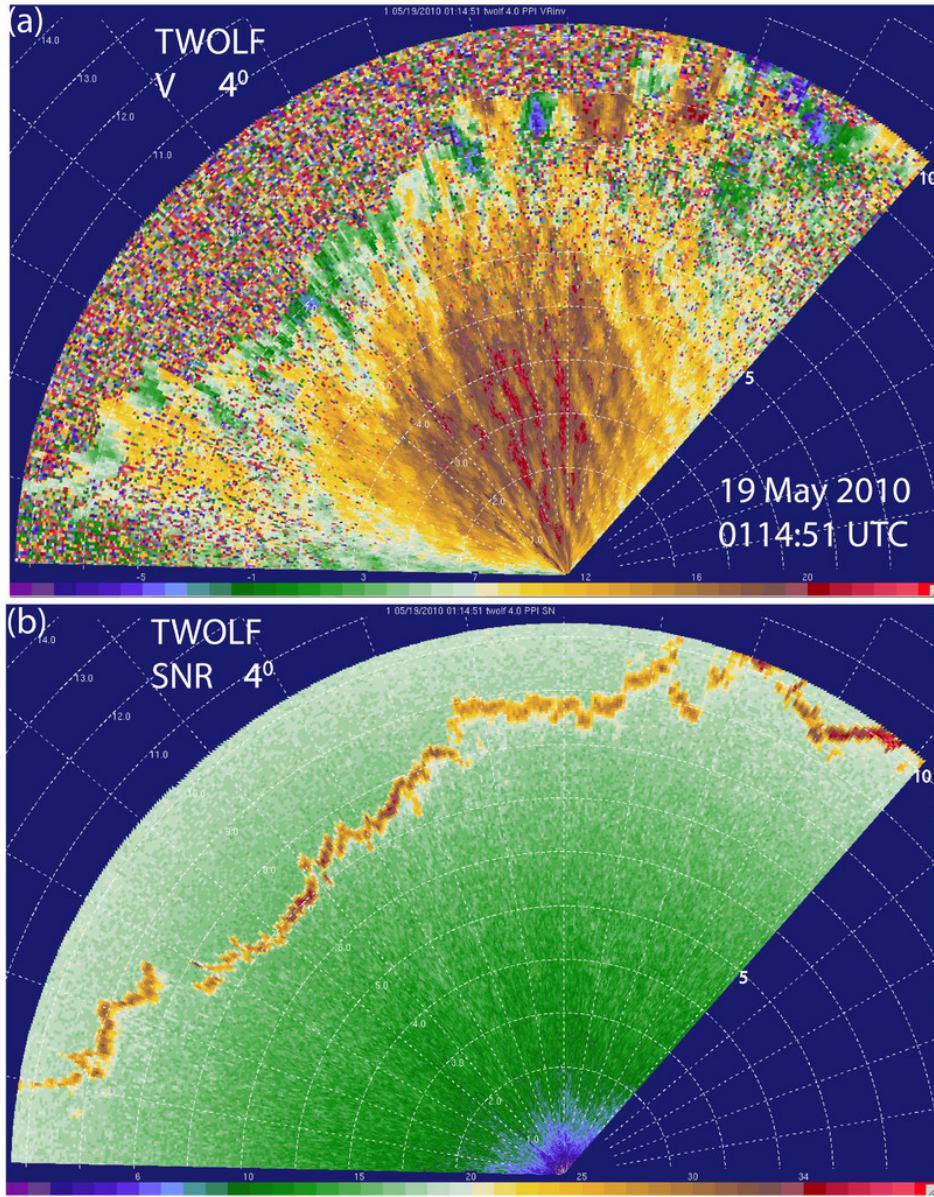


Figure 5.6: TWOLF PPIs of radial velocity (a) and signal-to-noise ratio (SNR, b) at 4.0° elevation from Fig. 7 in Bluestein et al. (2014).

case is uncertain, one could expect ranges $\approx 0\text{--}3$ km of this PPI to be in the ASL, and ranges > 3 km to be in the mixed layer of the PBL. Along the gust front near the edge of the observable lidar range, alternating positive and negative radial velocities are seen, similar to Palmer et al. (2011). These would be indicative of vortices along the boundary.

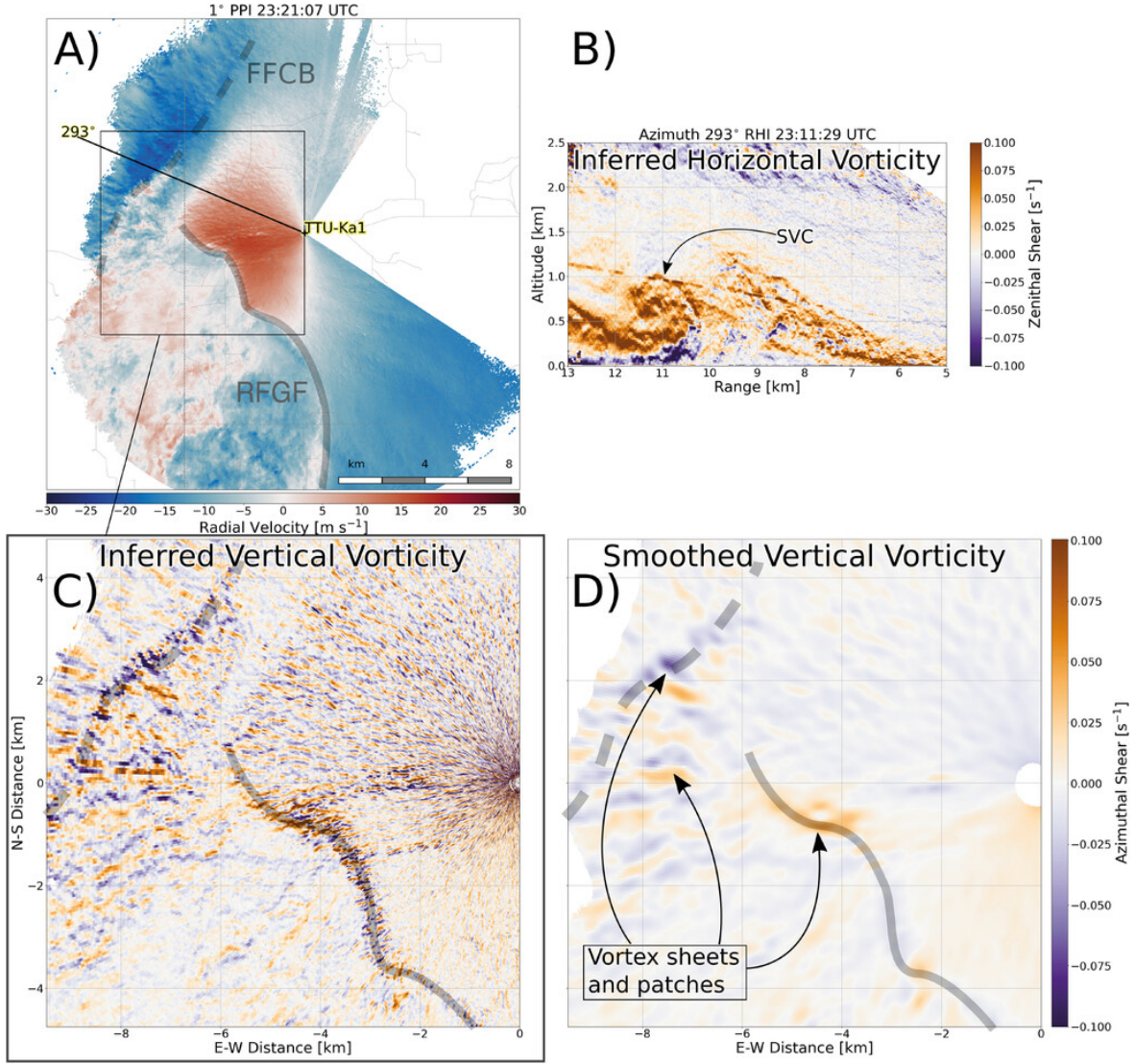


Figure 5.7: PPIs from the TTU mobile Ka-band radar of radial velocity (a), inferred horizontal vorticity through the SVC (b), inferred vertical vorticity (c), and smoothed vertical inferred vertical vorticity (d) from Fig. 3 in Fischer et al. (2024).

Mobile Ka-band radar observations by Texas Tech University (TTU, originally introduced by Schueth et al. (2021)) are shown in Fischer et al. (2024). Their PPIs of radial velocity, inferred horizontal vorticity through a streamwise vorticity current (SVC), and inferred vertical vorticity along the FFCB and RFGF are presented in Fig. 5.7. Because the beamwidth of the TTU Ka radars is 0.3° they are able to resolve larger magnitudes of ζ_{inf} because of the smaller azimuthal spacing and cross beam

distance. In Fig. 5.7, vorticity worms are seen in the inflow east of the FFCB and RFGF, with ζ_{inf} magnitudes of 0.1 s^{-1} in some structures. Along the FFCB and west of the RFGF, larger wavelength patches of magnitude $|\zeta_{inf}| > 0.1 \text{ s}^{-1}$ are present. Additionally, smoothed vertical vorticity (Fig. 5.7d) highlights a larger region of positive ζ_{inf} along the RFGF, similar to the structures seen in storm I1 on 10 May 2010.

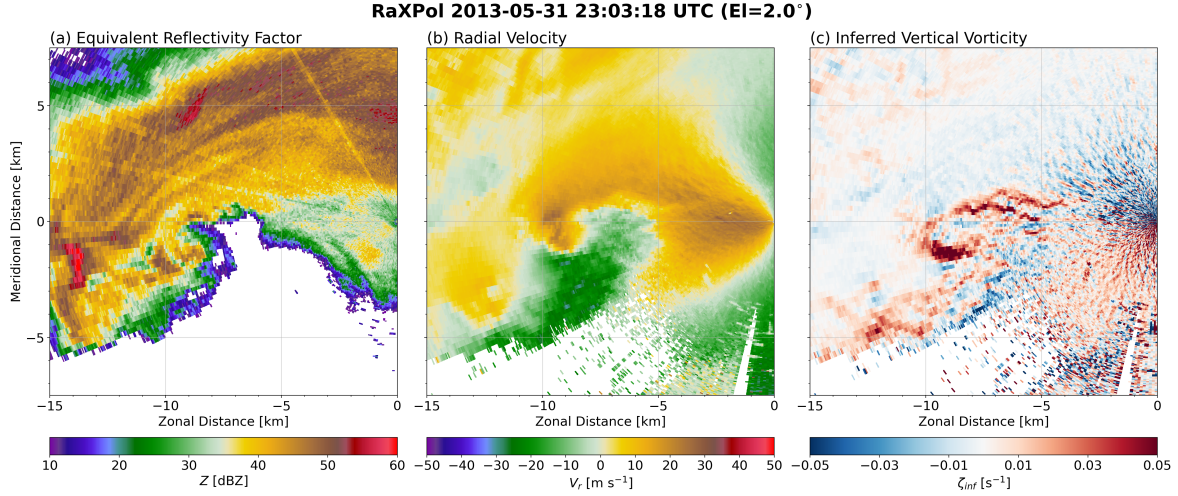


Figure 5.8: RaXPol 2.0° elevation PPI of reflectivity (a), radial velocity (b), and inferred vertical vorticity (c) approximately at the time of the formation of the 31 May 2013 El Reno, OK tornado.

Another example of inflow vorticity worms is in observations by the Rapid X-band Polarimetric (RaXPol) radar of the 31 May 2013 El Reno, Oklahoma tornado, which was the widest tornado ever recorded (Bluestein et al., 2015, 2019). RaXPol PPIs of reflectivity, radial velocity, and ζ_{inf} are presented in Fig. 5.8 approximately at the time of tornadogenesis. The vortex associated with the tornado is seen near $x = -9, y = -1.5 \text{ km}$ behind the RFGF, which extends to the east then of the tornado then arcs south. Low-speed streaks associated with vorticity worms are seen in the inflow east of the RFGF (between $x = -5$ and 0 km), and two distinct elongated vortices are seen along the RFGF east of the tornado. Similar to many of the previous cases analyzed herein, vortex patches are seen behind the RFGF in close proximity to

the tornado.

The plethora of examples of vorticity worms in supercell inflow in the existing literature shows that their presence is not unique to storm I1 herein or the Markowski (2024) simulations. Furthermore, they highlight that vorticity worms can exist in the inflow of supercells across different environmental conditions. Further analysis of the cases presented herein will provide better detail on the amplification of vorticity worms along gust fronts and underneath mesocyclones. Temporal resolution of < 60 s is likely needed to coarsely document the presence of a vortex merger, while much finer temporal resolution would better characterize the actual merging process (10–30 s). For example, past phased array have documented vortex mergers using 5–10 volumetric data (e.g., French et al., 2015; Kurdzo et al., 2017). This analysis would also provide better insight into whether vorticity worms facilitate or act as seeds for tornadogenesis in real supercells.

Chapter 6

Conclusions

The finding that vorticity worms facilitate supercell tornadogenesis in the Markowski (2024) simulations has yet to be directly corroborated in observed storms. Gebauer et al. (2026) presented Doppler lidar observations of vorticity worms in supercell inflow, finding interactions between mesocyclones, gust fronts, and inflow vorticity. However, these supercells were all non or pre-tornadic, preventing them from commenting on the ability of vorticity worms to instigate the tornadogenesis process. This study provides the first targeted analysis of vorticity worms during tornadogenesis in an observed supercell. The high-sensitivity OU-PRIME observations on 10 May 2010 provide sufficient radial velocity estimates below 100 m AGL in all regions of storm I1, allowing us to characterize vertical vorticity during tornadogenesis.

The OU-PRIME radial velocity observations reveal extreme winds ($> 40 \text{ m s}^{-1}$) near the ground and vorticity worms with $|\zeta_{inf}| > 0.05 \text{ s}^{-1}$ shortly before and during tornadogenesis. Very near OU-PRIME, we estimate the wavelength of ζ_{inf} by counting the positive vorticity worms that intersect a 3-km wide line orthogonal to the flow. Using this, we find a gradual increase in vorticity worm wavelength with height, increasing from ≈ 200 to 800 m in the lowest 300 m AGL. Following tornadogenesis, there is generally a decrease in wavelength at all heights analyzed, which coincides with a decrease in the inflow wind speeds near OU-PRIME. PPIs of ζ_{inf} wavelet power at 0.25° elevation quantitatively support the existence of vorticity worms in the inflow, and non-preferentially oriented vortex patches in the RFD 3 minutes before tornadoge-

nesis. At 2.0° elevation, minimal ζ_{inf} power is seen in the inflow, likely due to the radar beam overshooting the shallow vorticity worms. Similar magnitude wavelet power is seen in the RFD at 2.0° as at 0.25° elevation, indicating variance in the ζ_{inf} field is present at higher altitudes than in the inflow.

Partitioned time series of summed ζ_{inf} wavelet power in the inflow and outflow are presented to characterize the spatial scale of vertical vorticity within a 4-km radius of the mesocyclone. At 0.25° elevation, wavelengths $200 < \lambda < 800$ m are the dominant spatial scale of ζ_{inf} in both the inflow and outflow for most times. Power is larger in the outflow partition due to larger variance in this region. The wavelet time series show that ζ_{inf} in the inflow becomes more north-south oriented 5 minutes before tornadogenesis, which visually aligns with the 0.25° elevation PPIs. In the outflow, ζ_{inf} structures also appear to become more north-south oriented before tornadogenesis, but this may be due to the tornado vortex itself being partially located behind the RFGF and vortex patches along a secondary RFGF.

Using PPIs of ζ_{inf} and storm-relative winds from DDA, we comment on the vertical vorticity in the inflow and outflow of storm I1. Time-height series of circulation from DDA reveal similar values between the inflow and outflow in the lowest 1-km AGL. The southeasterly storm-relative winds in the inflow exceed 50 m s^{-1} before and during tornadogenesis while they are near zero and easterly in the RFD. We speculate that the inflow storm-relative winds are better oriented to advect vorticity towards the low-level mesocyclone than in the RFD. This discussion is corroborated by time-height series of vertical vorticity flux into the inflow and outflow partitions, which reveals entirely positive values in the inflow partition and negative values in the outflow. It is important to note this finding only provides an instantaneous description of vertical vorticity flux at the times in our DDA, which is limited by the temporal updates of the OU-PRIME and KTLX low elevation sweeps. Individual vorticity worms are identified on

four OU-PRIME 0.25° sweeps before tornadogenesis, and we comment on the available circulation and stretching potential in each. Due to the large area of vorticity worms ($> 10^6 \text{ m}^2$), circulation values exceed $10^4 \text{ m}^2 \text{ s}^{-1}$ for all times analyzed. Furthermore, we speculate the immense stretching potential of each vorticity worm may be large enough to drive pressure falls at the surface when subject to the stretching of the low-level mesocyclone and initiate stage 3 of the Fischer et al. (2024) tornadogenesis model. If this were the case, this would indicate that vorticity worms can act as seeds for tornadogenesis, providing the initial pressure fall at the surface which can converge near-ground vorticity from all storm regions.

In addition to the vertical vorticity present in the inflow of storm I1, vertical profiles of wind speed constructed from Oklahoma Mesonet observations and OU-PRIME radial velocity estimates at different elevations reveal a shallow layer ($\approx 50 \text{ m}$ deep) of horizontal vorticity exceeding 0.2 s^{-1} before tornadogenesis. Comparing the angle between the storm-relative and ground-relative shows substantial streamwiseness in the inflow, indicating this large magnitude horizontal vorticity near the ground may be properly oriented to be tilted beneath the mesocyclone and contribute to a surface vortex. This finding is similar to the observations in Gebauer et al. (2026) and simulations performed by Schenkman et al. (2014), which highlight frictional generation of horizontal vorticity in supercell inflow.

These findings motivate discussion of the angle between storm-relative winds and the orientation of the RFGF. When storm-relative winds in the inflow have a large easterly component, vorticity in the inflow cannot easily pass beneath the mesocyclone and is simply advected south and westward down the gust front. Under this scenario, near ground vertical vorticity must originate within the storm’s downdraft region, as that is the only region where the storm-relative winds point towards the mesocyclone. In contrast, when inflow storm-relative winds have a larger southerly component and

the RFGF does not surge far ahead of the mesocyclone, inflow vorticity is advected towards the mesocyclone and can more easily be amplified beneath it. This would indicate that strong rightward deviations in storm motion could enhance the pathway for inflow vorticity to facilitate tornadogenesis. This is particularly interesting given the findings of Homeyer et al. (2025), which found that the tornadic potential of observed right moving supercells is sensitive to the speed and direction of storm-motion deviations. Furthermore, Muehr et al. (2024) highlights the formation of horizontal rotors along the southern flank of mid-level mesocyclones when mid-level shear and low-level storm-relative flow are large, which may drive rightward deviations in storm motion. This speculation highlights a need to better characterize which environmental conditions control this angle between inflow storm-relative winds and the RFGF orientation. Furthermore, the atmospheric conditions that influence the presence, magnitude, and coherence of vorticity worms is not fully understood. Analysis of the simulations run in Frost et al. (2025) confirm that larger geostrophic winds generally support larger magnitudes of horizontal and vertical vorticity in the surface layer, but more analysis is needed to see how different surface heat fluxes influence the “streakiness” of vorticity worms. The University of Oklahoma Advanced Radar Research has operated a MetroWeather Doppler lidar collocated with twice daily radiosonde launches for much of the last year, performing alternating VADs and low-elevation PPIs to observe turbulent coherent structures in the surface-layer. This long period of data collection provides an opportunity to better characterize the nature of vorticity worms under varying environmental conditions.

In addition, it is unclear how the coherence of vertical vorticity at the ground impacts tornadogenesis. Parker (2023) discusses that larger wavelength vertical vorticity structures can provide more positive circulation, though given constant initial circulation, increasing the wavelength of vertical vorticity did not substantially increase the

likelihood of tornado formation. We do note that the Parker (2023) simulations only initialize quasi-random vertical vorticity. As shown herein, turbulent boundary layers possess substantial vertical and horizontal vorticity near the ground, meaning both tilting and stretching are important to surface vortex development. Given this, future studies could investigate the sensitivities of tornado formation to different boundary layer realizations consisting of differently sized and magnitude vorticity worms.

The findings of Gebauer et al. (2026) and our discussion of other supercell observations confirm that vorticity worms do exist in the inflow of other supercells and are not unique to the Markowski (2024) simulations or storm I1 on 10 May 2010. However, it is this author’s opinion that the tornadogenesis mechanism presented in Markowski (2024) cannot be directly confirmed with existing observations of supercells. Many past radar observations do not feature high enough sensitivity to retrieve accurate velocity estimations in the clear-air inflow of supercells. Additionally, many previous dual-Doppler observations of tornadogenesis used slower scanning radars and thus, have temporal updates on the order of multiple minutes, which have a similar inability to track the amplification of individual vortices as in our study. Rapid-scan radar observations and dual-Doppler set ups would be needed to achieve the temporal resolution required to perform backwards trajectory and circuit analyses, which would provide better insight into if vorticity worms facilitate tornadogenesis in real supercells. This is, however, complicated by the difficulty of clear-air retrievals by traditional radar wavelengths, such as S, C, and X-band. Doppler lidar has proven to be capable of observing vorticity worms very near the surface due to its narrow beam width and improved sensitivity in clear-air (Bluestein et al., 2014; Gebauer et al., 2026). However, Doppler lidar systems must scan slow (often $1\text{--}5^\circ \text{ s}^{-1}$) to accurately measure radial velocity, limiting the volumetric capabilities of these systems. Ka-band radars have also been successful at observing the clear-air inflow of supercells (Schueth et al., 2021).

Radar systems like the the Dual-Doppler 3D Mobile Ka-band Rapid-Scanning Volume Imaging Radars (KaRVIR; Salazar-Cerreno et al., 2022) in development at the University of Oklahoma could provide a unique opportunity to analyze vertical vorticity near the ground in supercells and provide the temporal resolution needed with its imaging capabilities.

Regardless of the challenges of confirming the Markowski (2024) tornadogenesis mechanism in real storms, further investigating the environmental conditions related to vorticity worms, storm-relative inflow and gust front orientation, and low-level mesocyclone strength will provide tornado forecasters with better understanding of tornadogenesis potential and potentially enable new prediction capabilities. Despite the advancements in the understanding of tornado formation that last few decades, it can still be very difficult to predict when and where tornadoes will form. Further advancing our understanding of the intricate dynamics of tornadogenesis and improving our communication methods could better equip forecasters, public safety officials, and the public to prepare for the impacts of tornadoes in the future.

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