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CHARACTERIZING THE EFFECTS OF LAND COVER CHANGES ON THE
SPATIOTEMPORAL AND MICROPHYSICAL PROPERTIES OF CONVECTIVE
CELLS IN HOUSTON, TEXAS

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CELLS IN HOUSTON, TEXAS

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Abstract

This study examines the behavior and microphysical properties of convective cells across different environments in the Houston metropolitan area during summer 2022. Using KHGX NEXRAD radar observations of 22,372 cell tracks, convective properties were analyzed across upwind, urban, and downwind regions.

The upwind region, dominated by coastal processes, exhibited peak cell initiation during early morning hours (80% of total cells) with smaller (42.5 km^2), less intense (36 dBZ) cells. The urban region showed approximately double the number of initiated cells per unit land area compared to upwind and 4–5 times more than downwind regions. Urban-influenced cells displayed greater variability in hydrometeor characteristics, larger mean raindrop sizes ($Z_{DR} \sim 1.8 \text{ dB}$), and higher liquid water content ($K_{DP} \sim 1.4 \text{ deg/km}$). Despite more frequent initiation, urban cells had lower precipitation efficiency. The downwind region demonstrated the highest precipitation efficiency (30 mm/cell) despite lower cell density.

Analysis of cell lifecycles revealed distinct patterns across regions. Urban impacts persisted downstream, with cells maintaining higher intensities through mature and dissipating stages. Upwind cells showed more uniform microphysical characteristics throughout their lifecycle, while downwind cells exhibited extended vertical development of ice particles during the developing and mature stages.

This research advances understanding of how urban environments alter precipitation formation mechanisms and how the combined effects of sea breeze and urban heat island influence the spatiotemporal evolution of convective storms. It also highlights the need for further investigation into the combined effects of sea breeze and urban heat island on convective storm evolution.

Chapter 1

Introduction

The global population trajectory is becoming more urban-centric. The world’s urban population quadrupled between 1950 and 2018, and urban areas are expected to expand for decades to come (United Nations, 2019). The United Nations’ 2018 Revision of World Urbanization Prospects (United Nations, 2019) report highlighted several key findings underscoring the expected expansion of urban areas globally, including projections for Houston:

- “By mid-century, roughly two-thirds (68%) of the world’s population will be living in urban areas. . . The global urban population is projected to grow by 2.5 billion urban dwellers between 2018 and 2050. . .”
- “North America is the region with the highest percentage urban and the second-to-last urban population size among all geographic regions. At the top, ten mega and large cities host 27% of the urban population, while another 29% lives in 41 medium-sized cities.”
- Houston’s population is expected to increase by 18.6% between 2018 and 2030.

Investigations into the potential impacts of increasing urbanization on the Earth system have revealed that cities can modify climate systems, leading to unique phenomena such as surface temperature increases (Chen and Frauenfeld, 2016; Qian et al., 2022), modifications in wind patterns and mesoscale circulations (Qian et al., 2022),

cloud enhancement (Qian et al., 2022), air pollution (Qian et al., 2022; Zhong et al., 2018), and storm and precipitation intensifications (Han et al., 2014; Qian et al., 2022).

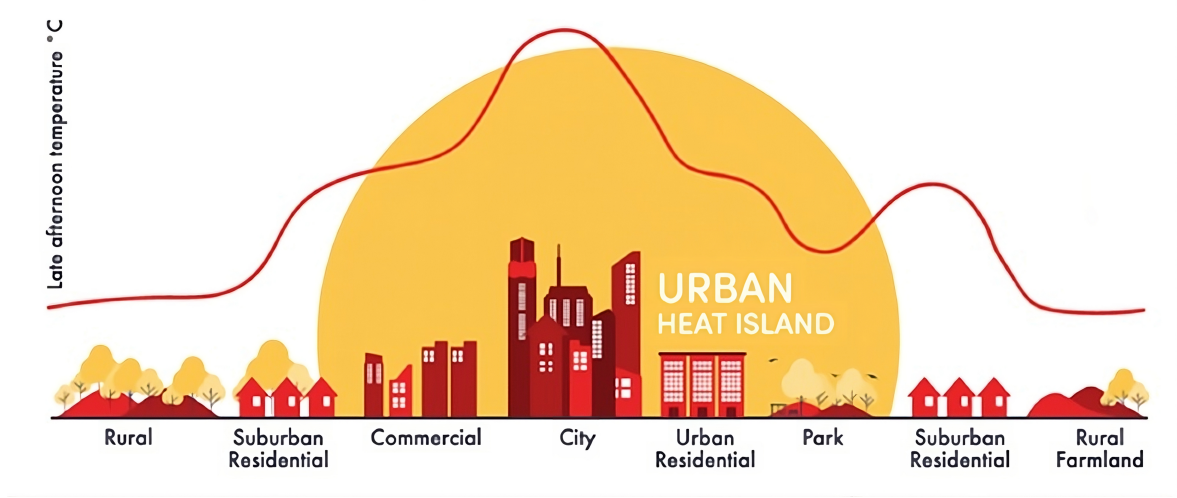


Figure 1.1: Schematic illustration of the urban heat island effect and temperature profile across an urban-rural landscape (Qian et al., 2022)

Much effort on understanding urban influences on local weather and climate has been devoted to the urban heat island (UHI) effect (Fig. 1.1), wherein urban areas exhibit higher temperatures relative to their rural surroundings (Deilami et al., 2018; Parker, 2010; Sarrat et al., 2006; Tam et al., 2015; Yow, 2007). This is pertinent, given that UHI has been found to have a considerable influence on mesoscale circulations and the resulting convection (Miao et al., 2011; Shepherd, 2005). Pioneering studies have revealed that urban areas and downwind regions exhibited rainfall increases of approximately 9-17% during the warmer seasons (Huff and Changnon, 1972). These early discoveries led to a more thorough investigation through the Metropolitan Meteorological Experiment (METROMEX) to understand how major urban areas affect mesoscale and convective precipitation processes. The key findings from the METROMEX studies pointed out a clear enhancement of rainfall in urban regions during the summer, with increases in precipitation ranging from 5% to 25% above the normal levels observed

within the cities and up to 50-75 km downwind (Changnon et al., 1971). Following this, numerous studies have validated and built upon these insights. For instance, Niyogi et al. (2011) analyzed 91 unique summertime thunderstorm cases in Indianapolis and found that storm characteristics were significantly altered by the urban region, especially storms near stationary-frontal boundaries. Additionally, Hand and Shepherd (2009) utilized ground-based and satellite-based datasets to investigate the spatial variability of warm-season rainfall events around Oklahoma City (OKC). Their analysis demonstrated that the downwind north-northeastern areas of the OKC metropolitan region experienced statistically significant increases in warm-season rainfall amounts relative to other surrounding areas. They concluded that the urban environment’s influence on precipitation modification was more substantial than the effects of agricultural land use or topographic features, particularly during weakly forced synoptic weather pattern regimes.

Urban areas can also profoundly influence convective storm systems due to their increased surface roughness relative to surrounding rural regions. This enhanced roughness, generated by buildings and infrastructure, can lead to greater low-level convergence over and downwind of cities, promoting uplift that can initiate convection (Rozoff et al., 2003). The resulting convective cells may intensify into severe thunderstorms (Bai et al., 2020, 2021), posing substantial threats to life, property, and economic activities through hazards like torrential rainfall, lightning, large hail, damaging straight-line winds, and tornadoes. Thus, an understanding of the behavior and characteristics of these cells is crucial for improving predictions and mitigating the hazards posed by urban-impacted severe convective storms.

Understanding convective storm dynamics and behavior is particularly critical for major coastal urban corridors like Houston. Coastal areas are prone to complex land-sea breeze interactions that can focus low-level convergence and uplift, priming the

environment for thunderstorm initiation (Bhate et al., 2016; Soderholm et al., 2016; Lu et al., 2006). In the Houston region, this land-sea breeze circulation is complicated by the convergence of the Gulf of Mexico sea breeze and the Galveston Bay breeze (Caicedo et al., 2019; Day et al., 2010; Fan et al., 2020; Shepherd and Burian, 2003). Houston’s low-lying coastal terrain and proximity to the Gulf of Mexico further increases its vulnerability to heavy rainfall and flood risks from these urban-influenced convective storms (Zhang et al., 2018).

To better understand the complex interactions between the coastal environment, urban surfaces, and convective processes in the Houston area, two major field campaigns were conducted: the Tracking Aerosol Convection interactions ExpeRiment (TRACER), from 1 October 2021 to 30 September 2022, and the Experiment of Sea Breeze Convection, Aerosols, Precipitation, and Environment (ESCAPE) campaign, from 30 May to 30 June 2022. The TRACER field campaign examined the influence of atmospheric aerosols on convective cloud processes in the urban environment. Its emphasis was on high-resolution tracking and comprehensive measurements of numerous individual, isolated convective cells throughout their entire lifecycle to establish linkages between the kinematic and microphysical evolution of convective clouds from initiation through dissipation. By employing an extensive suite of ground-based, airborne, and radar instrumentation, the campaign sought to quantify the influence of ambient environmental thermodynamic and kinematic conditions on the properties exhibited by convective cells during their lifecycle (Sheesley et al., 2023).

To further highlight the impact of sea breeze circulation in facilitating convective initiation and modulating precipitation patterns around the Houston area, the ESCAPE campaign utilized high-resolution radar and aircraft platforms to collect a comprehensive dataset of boundary layer observations, aerosol measurements, and cloud microphysics. The primary goals were to characterize convective cloud properties and

lifecycle under different aerosol and environmental conditions, quantify thermodynamic and kinematic controls on convective evolution, and improve process-level understanding of aerosol-convection interactions (Jensen et al., 2022).

A key insight from both campaigns is that the Houston metropolitan area introduces additional urban influences that likely modify sea breeze circulations. The urban landscape has an increased surface roughness compared to rural surroundings, which can enhance low-level convergence zones and impact heat fluxes, friction, and local circulations over the city (Chen et al., 2011). Additionally, the convergence of the UHI circulation with the bay and sea breeze patterns creates a complex preconvective environment that can destabilize the atmosphere and amplify forcing mechanisms conducive for thunderstorm development over and downwind of Houston (Shepherd and Burian, 2003). These observations were corroborated by the findings of Shepherd et al. (2010), who demonstrated through simulations that convective cells in the southwest and east of Houston are associated with advancing sea bay breeze front, and the UHI and urban surface characteristics of Houston can initiate and intensify convective storms. Their study revealed distinct rainfall maxima over and downwind of the urban area, attributed to the combined influences of surface fluxes, modified urban surface properties, and interaction with sea breeze circulation. Notably, they highlighted that convergence zones and a relative low pressure along the urban area edge were pivotal in intensifying convection, either alone or together with the sea breeze.

A critical aspect of convective storm systems is how the timing and location of convective initiation (CI) impact the morphology and characteristics of the resulting cells as they traverse different geographical areas (Byers and Braham, 1949; Fovell, 2005; Weckwerth and Parsons, 2006). Accurately anticipating when and where CI will occur is key, as the timing determines the initial environmental conditions present, while

the location sets the trajectory of the storm relative to transitioning surface boundaries. Although much progress has been made in understanding the overall structure and lifecycle of convective storms (Bodine and Rasmussen, 2017; Lombardo and Colle, 2010), discerning systematic changes in their characteristics due to underlying surface forcings remains a significant challenge, especially when factoring in boundaries between contrasting urban, rural, and suburban regimes and mesoscale circulations such as sea breezes (Liu and Xue, 2008). These circulation-induced perturbations are not fully understood nor adequately represented in numerical weather prediction models. Consequently, CI forecasts suffer from large uncertainties when convective systems transition between diverse surface regimes or mesoscale boundaries like sea breezes (Peters et al., 2019). Refining model parameterizations and improving predictions require deeper knowledge of how varying surface conditions and circulations from the upwind through downwind areas modify the precursor conditions, lifting mechanisms, and forcing features that facilitate convective storm development and initiation. Such comprehensive process-level understanding spanning the upwind environment to the urban region to the downwind area is critical for enhancing situational awareness and lead times related to CI events that can rapidly evolve into high-impact severe weather outbreaks across the varying surface regimes.

Meanwhile, some observational studies have highlighted the importance of understanding the near-cloud environment of CI processes (Lock and Houston, 2014; Nelson et al., 2021). For example, Marquis et al. (2023) analyzed data from the CACTI field campaign and determined that low-to-mid level atmospheric moisture content exhibited a strong relationship with the probability of successful CI events compared to cases where CI failed to occur. Their work implied that the entrainment of dry air into developing updrafts in the lower-to-middle troposphere can act as a significant

impediment to CI. Another study by Lombardo and Colle (2010) revealed that isolated cellular convection tends to initiate under limited quasi-geostrophic forcing and moderate instability levels, with an average most unstable CAPE around 973 J/kg. However, a significant knowledge gap remains in how these ambient environmental conditions evolve as convective cells traverse heterogeneous landscapes.

Observational analyses have demonstrated their utility in elucidating the impacts of urban areas on precipitation patterns, as exemplified by the groundbreaking insights from the METROMEX, ESCAPE, and TRACER campaigns. However, they are resource and computationally expensive. In addition, despite providing valuable ground-truth insights, observational datasets possess inherent limitations in their spatial and temporal coverage as well as the variables measured. For example, traditional surface meteorological observations lack the requisite details about urban morphology, land cover characteristics, and boundary layer properties that modulate urban impacts on convective processes (Qian et al., 2022). As a result, numerical modeling studies have been increasingly favored to simulate urban surface exchanges and their influences on convection initiation and storm evolution (Carrió et al., 2010; Chiu et al., 2022; Huang et al., 2022; Ryu et al., 2013).

While modeling efforts have allowed for an enhanced process-level understanding, the representation of physical processes in models, such as the interaction between urban surfaces and atmospheric phenomena, often involves empirical values and assumptions that may not be universally applicable or may be based on limited observational data (Huszár et al., 2020; Villalba-Pradas and Tapiador, 2022). More importantly, their credibility and refinement ultimately relies on the availability of comprehensive, high-quality observational data for rigorous evaluation across diverse environments (Qian et al., 2022).

Using high-resolution radar and satellite observations, this study seeks to advance our understanding of how urban areas systematically modulate the characteristics and environments of convective storms compared to their surrounding areas. Specifically, it aims to answer the following questions:

1. Are there discernible differences in the morphology and characteristics of convective cells in urban versus surrounding rural and suburban areas? Do these characteristics change systematically as the cells move into and out of the urban area due to modifications in the underlying surface forcing, including potential interactions with coastal phenomena such as sea breezes?
2. How do the near-storm environment of the cells change in the urban region compared to surrounding rural/suburban areas? Do urban surface properties preferentially favor or inhibit convective initiation through modulations of instability, moisture, or lift?

The subsequent chapters of this thesis are structured as follows. Chapter 2 establishes the background regarding convective cell formation, characteristics, and the mechanisms influencing their development and evolution. Chapter 3 provides a description of the study area and datasets employed, as well as a detailed description of the radar analysis methods used. Chapters 4 and 5 present the results and discussion of this research, highlighting the key findings related to urban impacts on convective initiation and storm morphology. Finally, Chapter 6 synthesizes the conclusions drawn from this study and outlines recommendations for future work.

Chapter 2

Related Works

2.1 Radar Background

Radar, an acronym for radio detection and ranging, is an active remote sensing system that uses electromagnetic waves to detect, locate, and characterize targets within its operational range (Skolnik, 2008). It consists of a transmitter, antenna, receiver, and signal processor. The transmitter generates high-power electromagnetic pulses, which are then emitted into the atmosphere via the antenna. As these pulses propagate through the atmosphere, they interact with various targets, including precipitation, biological scatterers, and inorganic particles such as dust or smoke (Doviak et al., 2006) (Fig. 2.1). The interaction between the radar's pulses and atmospheric particles is governed by scattering principles. When an electromagnetic wave interacts with a particle, the particle scatters the energy in various directions, behaving similarly to a miniature antenna (Batool et al., 2020). The characteristics of this scattered energy depend on several factors, including the size, shape, composition, and orientation of the target. For hydrometeors, the particle's composition, whether liquid or ice, significantly influences the scattering process through its refractive index or relative permittivity (Kumjian, 2013a; Fabry, 2018).

When these electromagnetic pulses encounter hydrometeors or other atmospheric particles, a portion of the energy is scattered back towards the radar. The antenna, operating in a dual transmit-receive mode, captures these reflected signals. The receiver

then amplifies these weak return signals, and the signal processor analyzes them to extract valuable meteorological information such as reflectivity, velocity, and in the case of dual-polarization systems, additional parameters like differential reflectivity and correlation coefficient (Kumjian, 2013a; Scott et al., 2001).

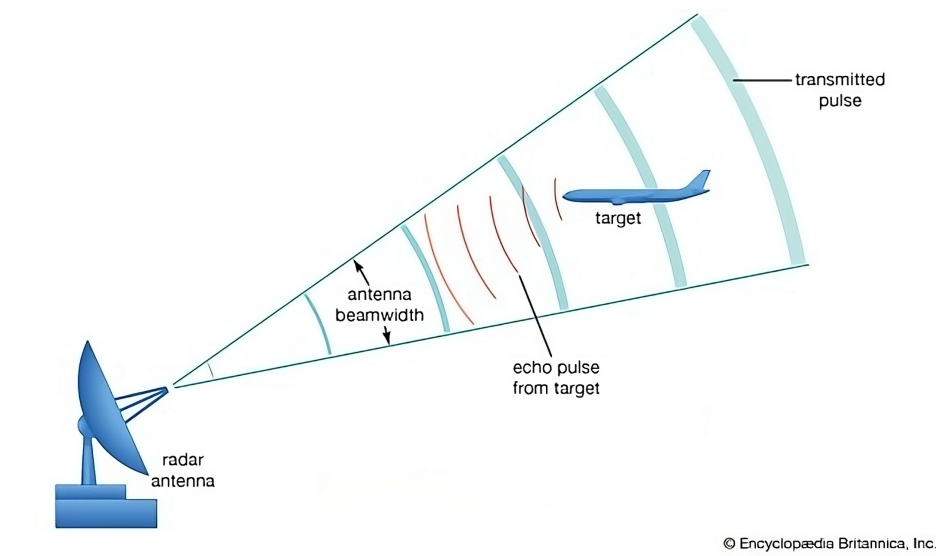


Figure 2.1: Image depicting basic radar operation¹

In conventional single-polarization weather radar systems, such as the early versions of the WSR-88D, the transmitted and received electromagnetic waves are horizontally polarized, meaning that the electric field vector oscillates in a plane parallel to the Earth's surface. However, modern dual-polarization radars transmit and receive both horizontally and vertically polarized waves, providing additional information about the shape, size, and orientation of hydrometeors. This allows for better precipitation type classification and improved rainfall estimates (Kumjian, 2013a; Scott et al., 2001).

¹<https://www.britannica.com/technology/radar>

The fundamental relationship between the transmitted and received power for a point target in weather radar systems is described by the equation (Doviak et al., 2006):

$$P_r = \frac{P_t G^2 \lambda^2 \sigma_b f^4(\theta, \phi)}{(4\pi)^3 r^4 L^2} \quad (2.1)$$

where P_r is the received power, P_t is the transmitted power, G is the antenna gain, and λ is the wavelength of the radar signal. The term σ_b represents the backscattering cross-section, while $f^4(\theta, \phi)$ is the two-way normalized antenna pattern, which depends on the elevation angle θ and the azimuth angle ϕ . The variable r denotes the range to the target, and L accounts for the one-way attenuation factor, often represented as losses. This equation is fundamental to understanding how weather radars detect and measure precipitation. It shows that the received power (P_r) is directly proportional to the backscattering cross-section (σ_b) and inversely proportional to the fourth power of the distance (r^4).

Radar systems use this principle to scan three-dimensional volumes of the atmosphere, and the radar scanning geometry is defined by three primary dimensions: range (r), azimuth (θ), and elevation (ϕ). Range represents the distance from the radar to the target measured along the path of the radar beam (Skolnik, 2008). It is calculated using the time delay (Δt) between pulse transmission and echo reception (Rauber and Nesbitt, 2018):

$$r = \frac{c \cdot \Delta t}{2} \quad (2.2)$$

where c is the speed of light. The factor of 2 accounts for the round-trip travel of the pulse. Azimuth (θ) is the horizontal angle of the radar beam, typically measured clockwise from the true north, while elevation (ϕ) is the vertical angle measured from the horizontal plane. By scanning in azimuth and elevation, the radar can scan a three-dimensional volume (Rauber and Nesbitt, 2018).

A common scanning strategy is the plan position indicator (PPI) scan, where ϕ is fixed and θ rotates through 360° . This creates a conical slice through the atmosphere, providing a quasi-horizontal cross-section of reflectivity or other radar variables at a constant elevation angle. Another strategy is the range height indicator (RHI) scan, where θ is fixed and ϕ varies, producing a vertical cross-section along a specific azimuth (Rauber and Nesbitt, 2018). Meanwhile, the WSR-88D radar employs volume coverage patterns (VCPs) as its primary scanning strategy. VCPs are predefined sets of elevation angles and rotation rates designed to efficiently sample the atmosphere in three dimensions. The radar cycles through these elevation angles to complete a volume scan, with different VCPs optimized for various weather conditions. For example, VCP 11 uses 14-16 elevation angles for severe weather, while VCP 31 uses 5 angles for clear air conditions, balancing factors such as update frequency, vertical resolution, and sensitivity to different atmospheric phenomena (Rauber and Nesbitt, 2018; Brown et al., 2000).

The spatial resolution of radar measurements is a critical aspect of these scans and is controlled by two main factors: range resolution and angular resolution (determined by beamwidth). Range resolution (Δr) is the radar's ability to distinguish between two targets along the same radial path. It is primarily determined by the pulse duration τ :

$$\Delta r = \frac{c \cdot \tau}{2} \quad (2.3)$$

where c is the speed of light. This equation shows that shorter pulse durations result in finer range resolution. For example, a pulse duration of $1 \mu\text{s}$ yields a range resolution of about 150 m. However, shortening the pulse duration reduces the total energy transmitted, which can negatively impact the radar's sensitivity and maximum range. To mitigate this, many modern radars employ pulse compression techniques, which

allow for long pulses with good range resolution. Meanwhile, angular resolution is determined by the radar's beamwidth, which is a function of the antenna diameter D and wavelength λ :

$$\Theta = K \frac{\lambda}{D} \quad (2.4)$$

where Θ is the beamwidth in radians and K is the beamwidth factor. Smaller beamwidths provide better angular resolution. The WSR-88D, for instance, operates at S-band ($\lambda \approx 10$ cm) with a 28-foot diameter antenna, resulting in a beamwidth of about 0.95° (Rauber and Nesbitt, 2018).

Determining the height of the radar beam is crucial for accurately interpreting radar data, especially at longer ranges. The height of the radar beam above the Earth's surface is primarily a function of the radar's elevation angle, the range from the radar, and the curvature of the Earth. In a simplified flat-Earth approximation, the height of the beam centerline would be:

$$h = r \cdot \sin(\phi) \quad (2.5)$$

where h is the height, r is the range, and ϕ is the elevation angle. However, this approximation becomes increasingly inaccurate at longer ranges due to Earth's curvature. To account for Earth's curvature and atmospheric refraction, we use a more comprehensive equation (Rauber and Nesbitt, 2018):

$$h = \sqrt{r^2 + (k_e \cdot R_e)^2 + 2 \cdot r \cdot k_e \cdot R_e \cdot \sin(\phi)} - k_e \cdot R_e \quad (2.6)$$

where h is the height of the beam above the Earth's surface, r is the range from the radar, R_e is the radius of the Earth (approximately 6371 km), ϕ is the elevation angle of the radar beam, k_e is the effective Earth radius factor, typically about 4/3

for standard atmospheric refraction. The k_e factor accounts for the slight downward bending of radar beams due to atmospheric refraction. This 4/3 Earth radius model combines the effects of Earth’s curvature and typical atmospheric refraction into a single correction factor.

The range of observations significantly impacts the radar beam height in several ways (Hunter, 1996; Kurdzo and Palmer, 2012; Chu, 2002):

1. Non-linear height increase: As range increases, the beam height increases non-linearly due to both the elevation angle and Earth’s curvature. This can lead to overshooting of low-level meteorological phenomena at far ranges.
2. Beam broadening: The beam width increases with range, resulting in a larger vertical and horizontal extent of the beam.
3. Atmospheric layer sampling: At longer ranges, the radar beam may sample entirely different parts of the atmosphere compared to closer ranges. For instance, at an elevation angle of 0.5° and a range of 100 km, the beam could be over 1 km above the surface, potentially missing low-level features.
4. Increased uncertainty: Variations in atmospheric refraction become more significant at longer ranges, leading to greater uncertainty in the exact beam position.
5. Partial beam filling: At longer ranges, the beam may only partially intersect with meteorological targets, potentially leading to underestimation of intensity or mischaracterization of phenomena.

These effects can lead to the potential underestimation of low-level phenomena, apparent height differences in melting layer signatures, and challenges in comparing reflectivity or other variables at different ranges. To mitigate these issues, radar networks often employ multiple radars with overlapping coverage, and scanning strategies may

include additional low-elevation scans to better capture low-level phenomena (Kurdzo and Palmer, 2012; Hunter, 1996).

Range-dependent effects are crucial for the accurate interpretation of radar data. As the distance from the radar increases, several effects come into play, each impacting the quality and interpretation of radar measurements. The spatial resolution of the radar degrades with range due to beam spreading, resulting in larger pulse volumes at greater distances. This decrease in resolution can obscure small-scale features and lead to non-uniform beam filling, potentially underestimating the intensity of compact, high-reflectivity cores within storms.

Concurrently, higher range can lead to underestimation of near-surface precipitation and missed detection of important low-level features such as tornadoes or melting layers. The radar equation demonstrates that received power decreases with the square of the range (Eq. 2.1), reducing sensitivity to weak echoes at far distances. This power loss can make it challenging to detect light precipitation at long ranges, potentially biasing climatological studies or nowcasting applications.

Moreover, attenuation of the radar signal becomes increasingly significant at longer ranges, particularly for shorter wavelength radars. This attenuation can cause underestimation of precipitation intensity behind areas of heavy rainfall and, in extreme cases, complete signal loss (Hitschfeld and Marshall, 1954).

While advancements in signal processing and radar design have mitigated some of these challenges, the fundamental physical limitations remain. To address some of these limitations and extract more detailed information about precipitation and other meteorological phenomena, dual-polarization radar has been developed.

2.2 Dual-Polarization Measurement Principles and Interpretation of Radar Variables

Dual-polarization radar, also known as polarimetric radar, represents a significant advancement in weather radar technology due to its ability to transmit and receive electromagnetic pulses in both horizontal and vertical orientations simultaneously (Bringi and Chandrasekar, 2001) (Fig. 2.2). This approach contrasts with conventional single-polarization radar systems, which operate using only horizontally polarized waves. The dual-polarization technique provides information about the size, shape, and orientation of hydrometeors based on differences in the backscattered signals between the two polarizations, which are influenced by the physical properties of the scatterers (Kumjian et al., 2022). These differences are quantified through several key polarimetric variables, including differential reflectivity (Z_{DR}), co-polar correlation coefficient (ρ_{HV}), and specific differential phase (K_{DP}).

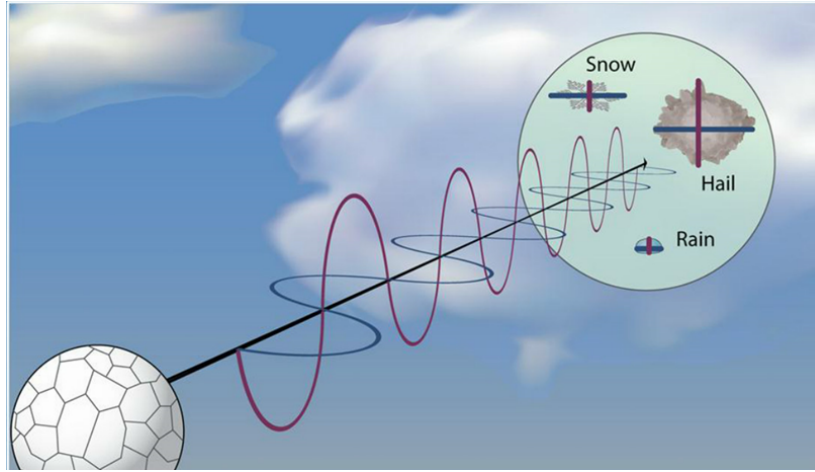


Figure 2.2: Illustration of the dual-polarization process²

²<https://www.weather.gov/news/130425-dualpol>

2.2.1 Radar Reflectivity Factor

Radar reflectivity factor (Z) is a measure of the power scattered back to the radar receiver from a specific sampling volume. Mathematically, it is the sum of the sixth powers of the particle diameters per unit volume, and it is given in terms of the hydrometeor size distribution by (Doviak et al., 2006):

$$Z = \int_0^{\infty} N(D) D^6 dD \quad (2.7)$$

where D represents the particle diameter and $N(D)$ denotes the drop size distribution. Z is typically expressed in units of $\text{mm}^6 \text{m}^{-3}$; however, for practical applications, it is often converted to a logarithmic scale as (Bringi and Chandrasekar, 2001):

$$Z(\text{dBZ}) = 10 \log_{10} \left(\frac{Z}{\text{mm}^6 \text{m}^{-3}} \right) \quad (2.8)$$

Z is directly related to the size and concentration of hydrometeors within the radar beam. It increases with increasing number density of particles, and it is significantly affected by the particle composition. For a given particle size, liquid hydrometeors reflect substantially more power than ice due to the higher dielectric constant of water. Mixed-phase particles containing both liquid and ice typically produce intermediate levels of backscattered power (Kumjian et al., 2022).

In the context of rainfall, Z generally ranges from 20 to 50 dBZ, with light rain typically having values below 30 dBZ, moderate rain having values within 30 and 40 dBZ, and heavy rain having values that exceed 45 dBZ (Fabry, 2018). These values are vital for estimating rainfall rates, often described through Z - R relationships of the form $Z = aR^b$, where R is the rainfall rate and a and b are empirically derived coefficients (Marshall and Palmer, 1948). However, these Z - R relationships can vary

significantly depending on the precipitation regime and drop size distribution (Battan, 1973). Meanwhile, for ice hydrometeors commonly found in convective storms, Z values can exhibit a broad range, with small ice crystals typically having Z values below 20 dBZ and aggregates or snowflakes falling within the 20-35 dBZ range. Additionally, graupel often exhibits values between 30 and 45 dBZ due to its higher density than snow, and hail, particularly large ones, can exhibit exceptionally high Z values that often exceed 55 dBZ (Kumjian and Ryzhkov, 2008).

Vertical profiles of Z within convective storms can provide valuable insights into microphysical footprints. However, Z for microphysical analysis in convective storms requires careful consideration, as it is most relevant to downdrafts. A sharp increase in Z below the melting layer may indicate collision-coalescence and rapid raindrop growth (Kumjian et al., 2022). Above the melting layer, an increase in Z toward the ground may indicate different ice microphysical processes, including vapor deposition, aggregation, riming, and melting (Kumjian et al., 2022). Similarly, a column of high Z extending well above the environmental freezing level (often termed a “reflectivity core”) can suggest the presence of a strong updraft capable of suspending large, potentially supercooled water drops or hail. These features are crucial for understanding storm structure and evolution (Kennedy et al., 2001).

Despite its utility, Z has limitations in fully characterizing the microphysical properties of hydrometeors. Although high Z values in the upper levels of a storm can indicate the presence of large ice particles such as hail or graupel, it alone cannot definitively distinguish between different types of hydrometeors or microphysical processes. For example, a region of moderate Z could represent either a high concentration of small particles or a lower concentration of larger particles, since Z is proportional to both the particle diameter and number concentration. Moreover, in mixed-phase regions of storms, interpreting Z becomes particularly challenging due to the different scattering

properties of coexisting liquid and ice hydrometeors. These limitations have led to the development and utilization of additional polarimetric variables, such as differential reflectivity, correlation coefficient, and specific differential phase. In combination with Z , these variables offer a more comprehensive understanding of hydrometeor types, sizes, and concentrations within convective storms.

2.2.2 Differential Reflectivity

Differential reflectivity (Z_{DR}) provides crucial information about the shape and orientation of hydrometeors within a radar sampling volume. It is the ratio of the reflectivity factors at horizontal and vertical polarizations and is typically expressed in logarithmic units. Mathematically, it is given by (Kumjian, 2013a):

$$Z_{DR} \text{ (dB)} = 10 \log_{10}(Z_h/Z_v) \quad (2.9)$$

where Z_h and Z_v are the reflectivity factors at horizontal and vertical polarizations, respectively.

The physical basis of Z_{DR} lies in the non-spherical nature of many hydrometeors. As raindrops fall below the melting layer, aerodynamic forces cause them to flatten into an oblate spheroidal shape. This shape results in a larger horizontal cross-section compared to the vertical, leading to stronger backscatter in the horizontal polarization (Kumjian, 2013a). Consequently, larger raindrops, which are more oblate, produce higher Z_{DR} values than smaller raindrops. Meanwhile, for hydrometeors that are small compared to the radar wavelength (Rayleigh scatterers), the orientation of their major axis determines the sign of Z_{DR} . Horizontally oriented particles produce positive Z_{DR} , while vertically oriented particles yield negative values.

Z_{DR} is also influenced by the physical composition of particles. For particles of the same size and shape, raindrops produce larger Z_{DR} values compared to ice owing to the higher refractive index of water. This property allows Z_{DR} to distinguish between different types of hydrometeors, including various forms of ice particles found in convective storms (Kumjian, 2013a).

For rain, Z_{DR} typically ranges from 0 to 4 dB. Small, nearly spherical droplets (< 1 mm diameter) produce Z_{DR} values close to 0 dB. As the drop size increases, Z_{DR} also increases, with values around 1 dB corresponding to median volume diameters of about 2 mm, and values exceeding 3 dB indicating the presence of large drops (> 5 mm diameter). This sensitivity to drop size allows Z_{DR} to be used in conjunction with Z to infer characteristics of the drop size distribution. However, for ice hydrometeors, the interpretation of Z_{DR} becomes more complex due to the various shapes, orientations, and densities involved, reflecting the diverse microphysical processes in the ice phase. Small, randomly oriented ice crystals produce Z_{DR} values near 0 dB due to their apparent spherical symmetry to the radar. Large, horizontally oriented ice crystals (such as dendrites or plates) can yield significantly positive Z_{DR} values, often in the range of 1-4 dB, due to their flattened shape (Kumjian, 2013a).

In convective storms, vertical profiles of Z_{DR} also provide insights into microphysical footprints. Below the melting layer where rain microphysical processes dominate, an increase in Z_{DR} is often associated with collision-coalescence, evaporation, and size sorting, whereas a decrease in Z_{DR} is often related to collisional breakup of larger rain particles into smaller ones. Above the melting layer, different ice processes such as aggregation, riming, and sublimation are often associated with a decrease in Z_{DR} , whereas an increase may be attributed to melting or vapor deposition onto pristine ice crystals (Kumjian et al., 2022).

A distinct feature in convective storms is the “ Z_{DR} column,” a layer of enhanced Z_{DR} extending above the environmental freezing level. This indicates the presence of supercooled liquid drops lofted by strong updrafts and is often associated with vigorous convection and potential hail formation (Kumjian, 2013b). The combination of Z and Z_{DR} also allows for improved hydrometeor classification. Regions of high Z but low Z_{DR} in the upper portions of a storm may indicate the presence of hail, while areas of moderate Z and high Z_{DR} likely contain large raindrops.

It is important to note that these microphysical footprints are typically interpreted by considering all polarimetric variables together, including ρ_{HV} and K_{DP} , to provide a more comprehensive understanding of storm microphysics. This helps to overcome some of the ambiguities inherent in using individual polarimetric variables alone and provides a more complete picture of the hydrometeor distribution and microphysical processes within convective storms.

2.2.3 Co-polar Correlation Coefficient

Copolar correlation coefficient (ρ_{hv}) provides valuable information about the diversity of hydrometeor types, shapes, and orientations within a radar sampling volume. It is a measure of the correlation between the horizontally and vertically polarized backscattered signals. Mathematically, it is given by (Doviak et al., 2006; Keat et al., 2016):

$$\rho_{hv} = \frac{\langle S_{VV} S_{HH}^* \rangle}{\sqrt{\langle |S_{HH}|^2 \rangle \langle |S_{VV}|^2 \rangle}}, \quad (2.10)$$

where $\langle S_{HH} \rangle$ and $\langle S_{VV} \rangle$ are the copolar elements of the backscattering matrix averaged over an ensemble of scatterers for the H and V polarizations, respectively, and the asterisk indicates the complex conjugate.

For a population of uniformly shaped and oriented particles, the backscattered signals at both horizontal and vertical polarizations will be highly correlated, resulting in ρ_{hv} values close to 1. Conversely, a mixture of different hydrometeor types, shapes, or orientations within the sampling volume will lead to decorrelation between the two polarizations, resulting in lower ρ_{hv} values (Kumjian, 2013a; Keat et al., 2016).

In the context of rain, ρ_{hv} typically exhibits very high values, generally above 0.98. This high correlation is due to the relatively uniform shape and orientation of raindrops. Slight variations in ρ_{hv} within rain can provide insights into drop size distributions and rainfall intensity. For instance, very heavy rain with a broad drop size distribution may show slightly lower ρ_{hv} values (around 0.95-0.98) due to increased diversity in drop shapes and possible turbulence-induced oscillations (Kumjian, 2013a; Bringi and Chandrasekar, 2001).

For ice hydrometeors commonly found in convective storms, ρ_{hv} values can vary significantly. Pristine ice crystals, such as needles and plates, can yield ρ_{hv} values less than 0.96. Graupel generally exhibits ρ_{hv} values similar to or slightly lower than dry snow, typically in the range of 0.95-0.98, depending on its density and the degree of riming (Kumjian, 2013a). Hail can produce a wide range of ρ_{hv} values, with dry, small hail having similar values to graupel (0.95-0.98) and large or wet hail having lower values ($\rho_{hv} < 0.85$) due to irregular shapes, tumbling motions, and water coating (Kumjian, 2013a).

In terms of storm microphysics, ρ_{hv} is often useful for identifying the melting layer signature, often referred to as “bright band” in radar observations (Matrosov et al., 2002). As snow begins to melt, the diversity of particle shapes and the presence of mixed-phase hydrometeors cause a significant decrease in ρ_{hv} , often to values between 0.8 and 0.9.

In the upper regions of convective storms, ρ_{hv} can help distinguish between different ice particle types. Regions of high ρ_{hv} coupled with moderate to high Z may indicate the presence of pristine ice crystals or small aggregates, while lower ρ_{hv} values in high-reflectivity areas could suggest the presence of large hail or mixed-phase precipitation (Kumjian, 2013a).

2.2.4 Specific Differential Phase

Specific differential phase (K_{DP}) provides crucial information about the concentration and shape of oriented hydrometeors within a radar sampling volume (Bringi and Chandrasekar, 2001). It is derived from the differential propagation phase (Φ_{DP}), which represents the cumulative phase difference between horizontally and vertically polarized waves as they traverse a medium containing anisotropic scatterers. This phase difference arises due to the interaction of electromagnetic waves with non-spherical particles, such as oblate raindrops. The oblate shape of particles causes horizontally polarized waves to slow down more compared to vertically polarized waves, resulting in a phase lag. This differential phase shift accumulates along the propagation path through the precipitation.

Mathematically, K_{DP} is defined as half the range derivative of Φ_{DP} (Kumjian, 2013a):

$$K_{DP} \text{ (}^\circ/\text{km)} = \frac{1}{2} \frac{d}{dr} \Phi_{DP} \quad (2.11)$$

where r denotes the range from the radar. This equation provides a measure of the differential phase shift per unit distance and is particularly useful for identifying regions of heavy precipitation and quantifying rainfall rates, especially in scenarios where traditional reflectivity-based estimates may be compromised (Kumjian, 2013a).

For rain, K_{DP} is particularly valuable due to its strong relationship with liquid water content and because it is not affected by the presence of hail. It typically ranges from 0 to 3 °/km for light to moderate rainfall, but can exceed 5 °/km in heavy rain (Ryzhkov et al., 2005). The relationship between K_{DP} and rainfall rate (R) is often expressed as:

$$R = a \cdot K_{DP}^b \quad (2.12)$$

where a and b are empirically derived coefficients (Ryzhkov et al., 2005). This relationship tends to be more robust than traditional Z-R relationships, especially in heavy rain, due to linear relationship of K_{DP} with liquid water content and its insensitivity to variations in drop size distribution (Bringi and Chandrasekar, 2001).

For ice hydrometeors, the interpretation of K_{DP} becomes more complex owing to the different shapes, orientations, and densities involved in ice-phase microphysics. Dry snow and small ice crystals generally produce K_{DP} values near 0 °/km due to their low dielectric constant and often random orientation. Large, horizontally oriented ice crystals (such as dendrites or plates) can yield positive K_{DP} values, typically in the range of 0.1 to 0.5 °/km, due to their shape anisotropy (Kennedy and Rutledge, 2011; Schneebeli et al., 2013). Graupel and small hail usually produce low K_{DP} values (< 0.2 °/km) due to their quasi-spherical shape and tumbling motions. In mixed-phase regions or areas with wet hail, K_{DP} can become notably positive (0.5 to 2 °/km) due to the presence of liquid water on the surface of ice particles (Ryzhkov et al., 2005).

In convective storms, the vertical profile and spatial distribution of K_{DP} provide valuable insights into microphysical processes and storm dynamics. Regions of high K_{DP} (> 2 °/km) often indicate areas of intense rainfall (Ryzhkov et al., 2005). Areas with moderate reflectivity but high K_{DP} may indicate the presence of large raindrops, which can be a precursor to hail formation. Columns of enhanced K_{DP} extending

above the environmental freezing level can indicate the presence of supercooled liquid water in strong updrafts (Kumjian and Ryzhkov, 2008). The combination of high reflectivity with low or negative K_{DP} can be indicative of hail, as large hailstones contribute significantly to reflectivity but little to K_{DP} due to their tumbling motion. Positive K_{DP} values in the upper levels of storms can indicate the presence of aligned ice crystals, providing information about electrification processes (Kumjian, 2013a).

Furthermore, increases in K_{DP} below the melting layer are usually associated with collision-coalescence, while decreases can indicate processes such as breakup, evaporation, and size sorting. Above the melting layer, increases in K_{DP} are related to vapor deposition or melting, while decreases suggest ice processes such as aggregation, riming, sublimation, or refreezing (Kumjian et al., 2022).

2.3 Advantages of Dual-Polarization Radar over Single-Polarization Radar

By leveraging the dual-polarization variables described above, dual-polarization technique offers several significant advantages over conventional single-polarization radar systems. One of the primary benefits of dual-polarization radar is its capacity for more accurate classification of hydrometeor types. By analyzing the shape and orientation information provided by polarimetric variables, we can better distinguish between rain, snow, hail, and mixed-phase precipitation (Straka et al., 2000). For example, for a S-band radar, rain can be identified by positive Z_{DR} , high ρ_{HV} , and positive K_{DP} in warm temperature regions; dry snow is characterized by near-zero Z_{DR} , high ρ_{HV} , and near-zero K_{DP} above the freezing level; and large hail can be identified by its high Z , near-zero or negative Z_{DR} , and lower ρ_{HV} values (Kumjian, 2013a).

Furthermore, dual-polarization radar can significantly improve the accuracy of quantitative precipitation estimation. Traditional single-polarization radars often struggle with errors in rainfall estimation due to factors such as partial beam blockage, attenuation, and calibration issues. However, the use of K_{DP} in rainfall estimation algorithms has been shown to mitigate these errors substantially (Ryzhkov et al., 2005). Even more significantly, dual-polarization enables the development of new rainfall estimation relationships that can eliminate the need for traditional Z-R relationships altogether. These dual-pol relationships often provide more accurate rainfall estimates across various precipitation types and intensities (Kumjian, 2013b).

Another significant advantage of dual-polarization radar is that it offers unprecedented insights into cloud and precipitation microphysics. For example, Z_{DR} and ρ_{HV} can be used to distinguish between pristine ice crystals and aggregates, while K_{DP} detects aligned ice crystals, indicating dendritic growth zones (Keat and Westbrook, 2017; Von Terzi et al., 2022). In the melting layer, the enhancement in Z_{DR} and reduction in ρ_{HV} allow can characterize the melting process, with K_{DP} changes indicating phase change rates (Kumjian and Ryzhkov, 2008). These detailed observations help identify microphysical processes such as ice formation, melting, and drop size distribution evolution, contributing to advancements in numerical weather prediction models.

2.4 Convective Cell Dynamics

Convective cells, specifically individual thunderstorms or clusters of thunderstorms, are a fundamental component of the Earth’s atmospheric system responsible for vertical heat and moisture transport, precipitation, and severe weather. Distinct from the broader global-scale convective circulations, such as the convection rising at the equator, convective cells in this study are characterized by the rising of warm, buoyant

air due to surface heating or other instabilities in the atmosphere. As the air rises, it undergoes adiabatic cooling based on the parcel theory (Doswell, 2001). When the rising air reaches the level of free convection, where it becomes less dense than the surrounding air, it continues to rise freely through a significant depth of the troposphere (Wallace and Hobbs, 2006). This process leads to the formation of convective clouds or storms (Fig. 2.1).

Convective cells can be broadly classified into three main types: single cell, multicell, and supercell thunderstorms (Houze, 2014). Single cell thunderstorms are isolated cells that typically have a short life span and limited intensity. Multicell thunderstorms, on the other hand, consist of multiple convective cells in various stages of development and decay, organized in clusters or lines. These include multicellular clusters and squall lines, which are a subset of multicell storms. Supercell thunderstorms are the most intense and organized type of convective cells, characterized by a deep, persistent rotating updraft known as a mesocyclone.

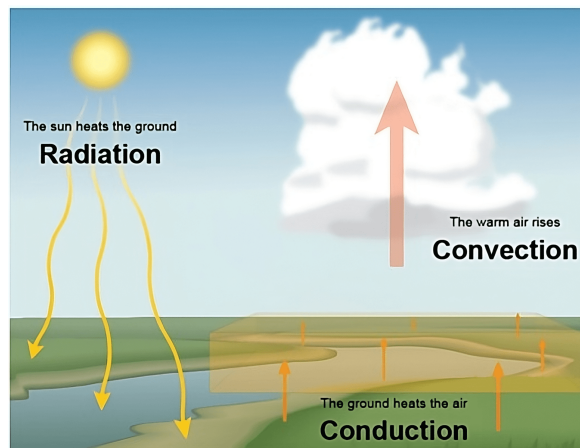


Figure 2.3: Illustration of the convection process³

In addition to this traditional classification, convective cells can also be classified into various types based on their depth (shallow or deep), width (wide or narrow),

³<https://www.thoughtco.com/what-is-convection-4041318>

and intensity (intense or weak) (Hong et al., 2024; Houze, 2014; Zipser and Liu, 2021). Shallow and deep cells involve vertical transport of heat and moisture across different depths. Shallow cells are confined within the boundary layer or lower troposphere and are initiated by surface heating, which creates an unstable layer near the ground (Henkes et al., 2021). They typically have a vertical extent of 1-3 km and are often associated with fair-weather cumulus clouds with limited vertical development and little to no precipitation. The vertical motions within these cells are driven by buoyancy forces and are generally weaker than those in deep convective cells (Tuftedal et al., 2023). Meanwhile, deep cells extend much higher into the atmosphere, often reaching the upper troposphere at 10-15 km altitude. These cells are fueled by strong buoyancy forces resulting from intense surface heating and higher convective available potential energy (CAPE), a measure of the amount of energy available for convection. They are also associated with towering cumulonimbus clouds, heavy precipitation, and sometimes severe weather.

Convective cells can also be classified based on their horizontal scale and intensity. Wide cells are characterized by their extensive horizontal extent, often spanning hundreds of kilometers (Houze, 2014). They can comprise multiple cells organized into complex structures, such as squall lines or mesoscale convective complexes (Fritsch and Forbes, 2001). The formation and maintenance of these cells are influenced by various factors, including large-scale atmospheric circulation, moisture availability, and wind shear (Rotunno et al., 1988). Intense wide convective cells, characterized by strong, sustained updrafts exceeding 10 m/s, are often associated with severe weather events such as heavy rainfall, damaging winds, and large hail (Gallus et al., 2008; Sánchez et al., 2003). These cells efficiently transport moisture and energy to higher levels of the atmosphere, fueling the development of towering cumulonimbus clouds and intense precipitation (Horváth and Geresdi, 2001). In contrast, narrow convective cells have a

more limited horizontal extent, typically ranging from a few kilometers to a few tens of kilometers, and are often isolated or loosely organized (Wilhelm et al., 2023). Narrow cells are more likely to develop in environments with weaker large-scale forcing and can be triggered by local instabilities or surface heterogeneities (Shusse et al., 2005).

Convection requires a mechanism to lift air to the level of free convection, and its initiation is influenced by different thermodynamic and dynamic factors (Jorgensen and Weckwerth, 2003; Weckwerth and Parsons, 2006). Thermodynamic instability, arising from surface heating and low-level moisture accumulation, allows air parcels to become buoyant and rise freely (Emanuel, 1994). However, the presence of thermodynamic instability alone does not guarantee convection initiation (Taszarek et al., 2021). Often, an additional lifting mechanism, known as dynamic forcing, is required to bring the air parcel to its level of free convection. Low-level convergence can provide dynamic forcing, occurring in the vicinity of mesoscale boundaries or through surface heterogeneities (Birch et al., 2014; Kalthoff et al., 2009; Kottmeier et al., 2008). One example of surface heterogeneity-induced convergence is the formation of sea breeze circulations. During the day, land surfaces warm up more quickly than adjacent water bodies due to their lower heat capacity. This differential heating creates a pressure gradient, with lower pressure over the land and higher pressure over the water. As a result, cooler air from the water flows towards the land at low levels, creating a sea breeze. Low-level air converges along the sea breeze and can lead to the development of convective clouds and precipitation along the coast (Allouche et al., 2023). Another form is uplift along boundaries, such as fronts, drylines, or outflow boundaries from previous convective storms (Fig. 2.2) (Intrieri et al., 1990; Weckwerth and Parsons, 2006). These boundaries often separate air masses with different densities and wind directions, leading to more persistent and vigorous low-level convergence that can trigger convection (Liu et al., 2023).

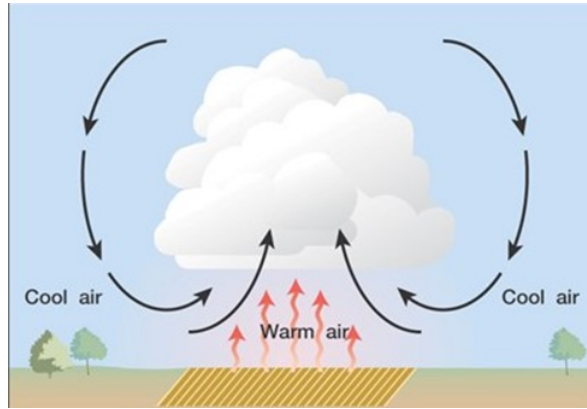


Figure 2.4: Lifting of air within a localized surface heterogeneity leading to the development of a cumulus cloud (Pecimotika, 2018)

The structure, evolution, and associated severe weather of a convective storm are linked to the characteristics of the storm’s environment (Lombardo and Colle, 2010; Mulholland et al., 2018). Factors such as atmospheric instability, wind shear, and moisture content have been found to play significant roles in determining the storm’s intensity, longevity, and structure. For example, Droegemeier and Wilhelmson (1985) employed a three-dimensional numerical model to study the effects of environmental wind shear on the evolution of simulated thunderstorms. Their results showed that the orientation of the updraft relative to the wind shear vector played a significant role in the storm’s structure and intensity. Storms that developed in environments with strong, unidirectional shear tended to have more erect and persistent updrafts, while those in weaker or more directionally varying shear exhibited more tilted and transient updrafts.

While the fundamental convective mechanisms are similar across the globe, the precise combination of factors that influence storm formation and evolution can vary substantially between regions. Land-sea contrasts, orography, and synoptic conditions can all modulate convective environments (Allouche et al., 2023; Tracy and Mecikalski,

2024). In Houston, for example, the convective environment is strongly influenced by coastal processes and the interactions between maritime and continental air masses (Statkewicz et al., 2021). The proximity of the Gulf of Mexico and the presence of the sea breeze circulation can lead to the development of convective storms along the coast, particularly during the summer months (Lombardo and Colle, 2010). The interaction between the sea breeze and the urban heat island effect can also modify the convective environment and influence the timing and location of storm initiation (Shepherd and Burian, 2003).

2.4.1 Environmental Conditions Influencing Convective Cell Development

The initiation and evolution of convective storms are highly sensitive to environmental conditions. A critical factor determining whether or not convection can initiate is the presence of a stable layer or lid of warm, dry air aloft, called convective inhibition (CIN). CIN is measured based on the work a parcel must do to overcome the stable layer and reach its level of free convection (Markowski, 2007). In some cases, the capping inversion allows high wet-bulb potential temperature air to accumulate in the boundary layer and increase instability prior to convective initiation (Browning et al., 2007). The strength, thickness, and spatial variability of the capping inversion can have a significant impact on where and when convection can break through (Browning et al., 2007; Abulimiti et al., 2023). Graziano and Carlson (1987) found that weak or spatially discontinuous lids are more favorable for widespread convective initiation, while strong, continuous lids can suppress convection until localized forcing, such as upper-level disturbances or surface convergence, is sufficient to overcome the inhibition. Temporal variations in lid strength, such as diurnal heating or advection of warm air

aloft, can also modulate convective potential. Additionally, the spatial variability of the capping inversion can create localized regions of enhanced instability, known as “hotspots,” where convection is more likely to initiate (Blanchard, 1998).

Boundary layer temperature and moisture are also key environmental factors that influence convective initiation, timing, and intensity. Studies have shown that even small changes in low-level temperature and moisture profiles can have significant impacts on whether or not storm development occurs (Crook, 1996; Fabry, 2006; McCaul et al., 2005). An increase in the low-level temperature lapse rate or a substantial increase in boundary layer mixing ratio can significantly enhance the potential for convective development. Conversely, a decrease in lapse rate or a drying of the boundary layer can inhibit convection. This sensitivity arises because boundary layer warming and moistening directly influence parcel buoyancy and equivalent potential temperature, which are key determinants of CAPE and CIN (Houston and Niyogi, 2007; Kirkpatrick et al., 2009). Stirling and Petch (2004) revealed that spatial variabilities in moisture and temperature can increase the onset of deep convection. By including model-based moisture and temperature variability in their experiments, they found that convection occurred 1-3 hours earlier and precipitation increased by 20-70%. Additionally, Bodine et al. (2010) utilized radar reflectivity data to investigate the influence of moisture variability on convection initiation. They found that the potential for convection initiation increased within a small-scale region of pooled moisture and convergence. This was characterized by a thermodynamic environment with near-zero CIN, a reduction in LFC from 2385 m to 1833 m, and considerably higher CAPE.

In addition to moisture variability and CAPE, wind shear also plays a critical role in storm dynamics, both in the boundary layer and aloft. Wind shear can have a significant impact on convective initiation by influencing the lifting of air parcels and the organization of resulting storms. In environments with strong vertical wind shear,

horizontal convective rolls can develop in the boundary layer, which can help to focus convergence and provide the necessary lift for convective initiation (Weckwerth and Parsons, 2006). Additionally, vertical wind shear can help to organize convection by promoting storm-scale rotation and updraft-downdraft separation (Rotunno et al., 1988). Once convection has initiated, the orientation of wind shear relative to storm inflow affects updraft strength, storm propagation, and organization (Fovell and Ogura, 1989). Strong deep-layer shear (for example, 0-6 km shear of 20-40 knots) generally promotes longer-lived and more intense convective storms by enhancing dynamic pressure forces and updraft rotation (Weisman and Rotunno, 2000). In this case, the wind shear can help to vertically tilt the updraft, preventing precipitation from falling back into the updraft core and reducing the negative effects of water loading and evaporative cooling. This allows the storm to maintain its intensity and organization for a longer period. However, excessively strong wind shear can also have a detrimental effect on storm development. If the shear is too strong, it can tilt the updraft to such an extent that it becomes detached from its source of warm, moist air at the surface. This can lead to a disruption of the updraft and eventual storm dissipation. Recent numerical simulations and observational studies have further investigated the role of environmental wind shear in controlling the intensity, structure, and longevity of convective storms (Fan et al., 2009; Frank and Ritchie, 2001). For example, numerical simulations by Weisman and Rotunno (2004) explored the influence of vertical wind shear on squall line evolution. They found that the orientation and strength of the ambient shear had a significant impact on the updraft intensity and overall system organization. Stronger deep-layer shear promoted more upright and intense updrafts, leading to longer-lived and well-organized squall lines.

2.5 Mechanisms of Urban Influence on Deep Convective Initiation and Evolution

Many studies have documented that urban areas exert a significant influence on the initiation, intensity, and spatial distribution of convective cells (Han et al., 2014; Qian et al., 2022; Rozoff et al., 2003). The unique physical properties of cities, including their high surface roughness, thermal characteristics, and moisture availability, modify local atmospheric conditions in ways that can favor or suppress convective activity. Several key mechanisms have been proposed to explain how cities influence convective cells and precipitation, including:

1. Enhanced surface roughness and low-level convergence: Urban structures, such as tall buildings and transportation infrastructure, significantly increase the aerodynamic roughness of the surface compared to natural landscapes. This enhanced roughness leads to increased mechanical turbulence and mixing in the urban boundary layer, as well as reduced wind speeds near the surface (Zhang et al., 2017). Consequently, low-level convergence is often enhanced downwind of cities, as faster-moving air aloft is forced to descend and converge with slower-moving air near the surface (Bornstein and Lin, 2000). This urban-induced convergence can provide the necessary uplift to initiate convection, particularly when coupled with favorable thermodynamic conditions.
2. Destabilization of the boundary layer from urban heat island effects: The UHI effect is a well-documented phenomenon in which cities exhibit higher temperatures than surrounding rural areas, particularly at night (El Afandi and Ismael, 2023). It arises from a combination of factors, including the high thermal conductivity and heat capacity of urban materials, anthropogenic heat sources (e.g., vehicles,

industry, air conditioning), and reduced evaporative cooling due to limited vegetation and pervious surfaces. This urban heating leads to the development of a thermal low-pressure system over the city, which can induce local circulation with rising motion over the urban center and subsidence in the surroundings (Vraciu, 2024). The UHI circulation, combined with the general warming and destabilization of the urban boundary layer, can create conditions favorable for deep convection, particularly in the presence of sufficient low-level moisture (Vraciu, 2024).

3. Cities as sources of moisture: While urban areas are often characterized as “moisture deserts” due to their extensive impervious surfaces and lack of vegetation, cities can actually serve as significant sources of atmospheric moisture (Kuttler et al., 2007). Anthropogenic activities, such as combustion, industrial processes, and air conditioning, release water vapor into the urban atmosphere (Mishra and Kannan, 2023). This enhanced moisture availability can influence convective processes by increasing the equivalent potential temperature and buoyancy of urban air parcels, as well as providing additional latent heat energy for storm intensification (Theeuwes et al., 2022).
4. Bifurcation or diversion of precipitating systems: The complex geometry and spatial inhomogeneity of urban surfaces can influence the movement and evolution of convective storms. The increased surface roughness and friction in cities can slow storm propagation and prolong the duration of heavy precipitation over urban areas (Shi et al., 2024). In some cases, the blocking effect of urban structures can cause storms to bifurcate and flank the city, leading to enhanced precipitation in the surrounding areas (Niyogi et al., 2011).

2.5.1 Enhanced surface roughness and low-level convergence

Several observational and modeling studies have demonstrated the impact of urban surface roughness on boundary layer processes and convective initiation (Fig. 2.3) (Duan and Takemi, 2021; Rajeswari et al., 2021; Shepherd et al., 2002). For example, Bornstein and Lin (2000) analyzed the data collected during the Project ATLANTA (ATlanta Land use Analysis: Temperature and Air quality) in Atlanta, Georgia and found a significant increase in the frequency and intensity of convective thunderstorms downwind of the city center. They attributed this enhancement to the convergence zone induced by the UHI effect and the background flow at different times of the day, which provided the necessary uplift to initiate convection. Numerical simulations conducted by Rozoff et al. (2003) provided further support for this finding, demonstrating that the increased surface roughness of the urban area led to enhanced low-level convergence and rising motion, which triggered convection when the atmosphere was sufficiently unstable. Similarly, Dixon and Mote (2003) expanded upon these findings and suggested that low-level moisture, in addition to the UHI intensity, significantly contributed to convection initiation and precipitation. Similar results have been reported for other regions, such as Paris, France (Thielen et al., 2000), and Germany (Kalthoff et al., 2009).

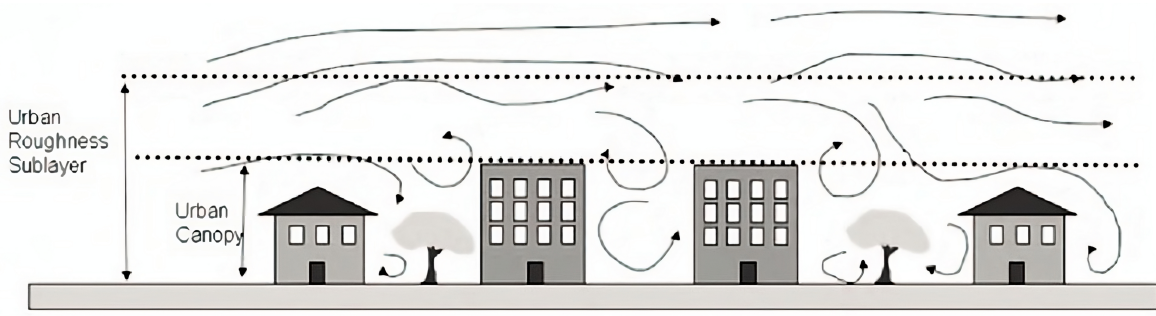


Figure 2.5: Urban surface roughness and boundary layer wind flow (Martilli et al., 2007)

The impact of urban surface roughness on convective processes is not limited to the initiation stage. Once storms form, the increased friction and drag over cities can influence their propagation and structure. Niyogi et al. (2011) used radar data to track the movement of thunderstorms over Indianapolis, Indiana. They found that storms tended to slow down and intensify as they approached the urban center, likely due to the enhanced convergence and upward motion induced by the roughness of the city. In some cases, storms split or bifurcated around the city, leading to prolonged periods of heavy rainfall in the surrounding suburbs.

2.5.2 Destabilization of the boundary layer from urban heat island effects

The UHI effect has long been recognized as a potential driver of urban-enhanced convection. The higher temperatures in cities compared to surrounding rural areas can lead to increased instability and rising motion in the urban boundary layer, which can trigger or intensify convective storms (Shepherd, 2005). Numerous studies have documented the impact of the UHI on convective precipitation (Ganeshan et al., 2013; Lin et al., 2011; Wu et al., 2019). Shepherd et al. (2010) analyzed rainfall data and

found that the UHI in Houston, Texas led to a significant increase in the frequency and intensity of summer thunderstorms over the city compared to the surrounding rural areas. They attributed this to the destabilization of the urban boundary layer by the UHI, which provided the necessary buoyancy for convective initiation. Similar findings were reported by Haberlie et al. (2015), who investigated the relationship between the UHI and convective precipitation in Georgia and Alabama. Their study revealed that the UHI-induced destabilization of the urban boundary layer increased the likelihood of convective initiation and the intensity of the resulting storms.

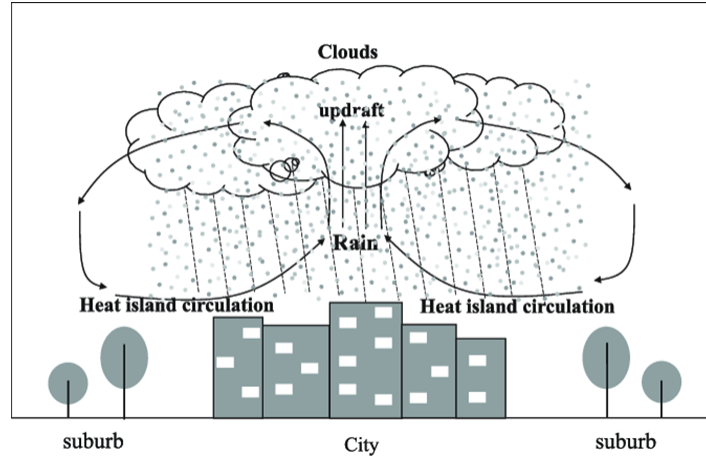


Figure 2.6: Simplified diagram of the urban heat island and precipitation (Yang et al., 2017)

The impact of UHI on convection depends on urban heating, background atmospheric conditions, and land surface properties. The UHI can enhance convection by increasing the instability and buoyancy of the urban boundary layer, which can lead to stronger updrafts and more intense storms (Fig. 2.4) (Chiu et al., 2022; Vraciu, 2024). However, the extent to which it influences convection varies depending on the prevailing atmospheric conditions. In situations where the atmosphere is moderately unstable and the background winds are weak, the UHI circulation can become more

pronounced and interact with the daytime heating, thus destabilizing the urban boundary layer (Meng et al., 2007; Zhong et al., 2015). This interaction can lead to enhanced convective activity over and downwind of the city.

Furthermore, the spatial pattern and intensity of UHI can also modulate its impact on convective processes. The UHI is not homogeneous across an urban area, but rather exhibits significant spatial variability depending on land use, building density, and surface properties (Zhang et al., 2022). In densely built-up areas, such as central business districts or industrial zones, the UHI effect tends to be most pronounced. These areas are characterized by a high concentration of impervious surfaces, such as buildings, roads, and parking lots, which have a low albedo and high heat capacity (Nguyen, 2023). These surface properties allow the urban fabric to absorb and store more solar radiation during the day, and then release this heat into the urban atmosphere at night. Meanwhile, areas with more vegetation and green spaces, such as parks, gardens, and suburban neighborhoods, tend to have a weaker UHI effect. The presence of vegetation can significantly modify the surface energy balance and the microclimate of these areas. Trees and other plants provide shade, reducing the amount of solar radiation reaching the ground. They also transpire water, which cools the air through latent heat absorption. As a result, these green areas can act as “cool islands” within the city, with lower surface and air temperatures compared to the surrounding built-up areas (Kraemer and Kabisch, 2022).

The spatial distribution of these cool islands can create a more heterogeneous thermal landscape within the city, with a mosaic of warmer and cooler patches. This heterogeneity can influence the location and strength of the UHI-induced convergence zones, and consequently, the patterns of convective activity over the urban area (Alahmad et al., 2020; Bornstein et al., 2012). For example, convergence zones may preferentially

form along the boundaries between the warmer built-up areas and the cooler green spaces, where the thermal gradients are strongest.

2.5.3 Moisture from cities

While the UHI and surface roughness effects have received the most attention in studies of urban convection, the role of cities as sources of atmospheric moisture is increasingly recognized as an important factor (Huang and Song, 2023; Mayer et al., 2003; Wang et al., 2021). Urban areas can influence local and regional moisture budgets through a variety of processes, including anthropogenic water vapor emissions, sea-breeze circulations, and changes in evapotranspiration from urban surfaces.

During summer, anthropogenic moisture fluxes from cities, especially industrial areas, can be substantial and significantly impact local humidity levels and boundary layer processes. These anthropogenic moisture emissions occur primarily through evaporative cooling and through chemical reactions during the combustion of hydrocarbon fuels (Mishra and Kannan, 2023). Gorski et al. (2015) estimated the emission of water vapor through fossil fuel combustion over Salt Lake City and found that it contributed up to 13% of the boundary layer vapor. While this moisture flux is relatively small compared to natural sources, it can have a significant impact on boundary layer processes, especially when natural evapotranspiration is minimal.

However, the net impact of cities on local moisture budgets depends on the balance between anthropogenic moisture sources and the reduction in evapotranspiration due to the prevalence of impervious surfaces in urban areas (Wang et al., 2021). In cities with extensive urban development and limited vegetation, the loss of evapotranspiration can potentially outweigh the anthropogenic moisture emissions. The relative importance of these factors may also vary depending on the climate and surrounding environment of the city. In arid and semi-arid regions, urban landscaping and vegetation are often

heavily irrigated to maintain greenness and mitigate the UHI effect, which can lead to significant evapotranspiration and moisture fluxes (Huang and Song, 2023).

The moisture released by cities can influence convective processes in several ways. First, it can increase the equivalent potential temperature and buoyancy of urban air parcels, making them more likely to rise and trigger convection (Wang et al., 2021). Second, it can provide additional latent heat energy for storm intensification and growth, particularly in the presence of strong updrafts and vertical moisture transport (Li et al., 2020). Finally, it can influence the microphysical properties of urban clouds and precipitation, such as the size and concentration of cloud droplets and ice crystals (Theeuwes et al., 2022; Wang et al., 2021).

2.5.4 Bifurcation or diversion of precipitating systems

In addition to their impact on convective initiation and intensification, cities can also influence the movement and structure of mature thunderstorms. The complex geometry and spatial inhomogeneity of urban surfaces can lead to the bifurcation, splitting, or diversion of storms as they approach or pass over cities. Compact-rise and high-density buildings in urban areas can act as physical barriers to the flow of air, creating a “building barrier effect.” When a storm approaches a city, the tall buildings can block or divert the flow, causing the storm to split or move around the urban core. This can lead to the concentration of precipitation on the flanks or downwind side of the city, as the storm cells converge and intensify in these areas (Dou et al., 2015; Han et al., 2014; Shi et al., 2024).

One of the earliest studies to document this effect was by Bornstein and LeRoy (1990), who analyzed radar data of thunderstorms approaching New York City. They found that storms often split or diverted around the city, with the most intense

precipitation falling in the suburbs to the north and east. They attributed this behavior to the increased surface roughness and friction of the urban area, which slowed the progress of the storms and caused them to bifurcate. Corroborating this finding, Niyogi et al. (2011) used radar data to track the movement of thunderstorms over Indianapolis. They found that storms often split or diverted around the city center, with the most intense precipitation falling in the suburban areas. They suggested that the tall buildings and increased surface roughness of the urban core created a barrier effect, causing the storms to split and intensify on the flanks of the city.

2.6 Impact of coastal urban areas and land-sea breeze interactions on deep convection

Coastal cities are particularly prone to urban-influenced convection due to the interactions between the urban boundary layer, sea breeze circulations, and the coastal atmospheric environment. As the land heats up during the day, it warms the air above it, creating a thermal low-pressure system. The cooler air over the sea then flows towards the land, forming the sea breeze. The strength and inland penetration of the sea breeze depend on various factors, such as land-sea temperature contrast, background wind flow, and coastal topography. The land-sea temperature contrast creates a pressure gradient that drives onshore flow during the day (sea breeze) and offshore flow at night (land breeze). These circulations can interact with the urban heat island and surface roughness effects to create regions of enhanced convergence and uplift, which can trigger convection if the atmosphere is sufficiently unstable (Shepherd et al., 2010).

Modeling studies have provided further insights into the mechanisms of coastal urban convection. Freitas et al. (2007) used a mesoscale model to simulate the impact of São Paulo, Brazil, on sea breeze convection. They found that the UHI circulation

interacted with the sea breeze to create a convergence zone over the city, which triggered deep convection when the atmosphere was sufficiently unstable. The storms that formed over the city were more intense and produced more lightning than those in the rural areas. Meanwhile, Hu et al. (2022) used the WRF model to simulate a moist convection event in Tianjin, China, caused by the interaction of the UHI and sea breeze front. They found that the heat island initially enhanced turbulence but weakened after precipitation. In addition, the urban surface obstructed the sea breeze front's movement, maintaining it near the city center and providing moisture for convection. These findings were validated by observational studies. For instance, Vemado and Pereira Filho (2015) used observational data to investigate the combined effects of the UHI and sea breeze on severe weather events in São Paulo. They found that the UHI circulation enhanced the sea breeze front and increased the low-level convergence over the city, leading to more intense and longer-lived storms. The combined effects of the UHI and sea breeze were shown to significantly influence the microphysical properties of precipitation, leading to heavier rainfall and more severe weather events in the São Paulo metropolitan area.

The impact of coastal cities on convection is not limited to sea breeze interactions. The increased surface roughness in coastal urban areas can reduce wind speeds and also alter the direction of the land-sea breeze circulation. Li et al. (2020) used the WRF model to study the impact of urbanization on the sea breeze in the Pearl River Delta region of China. They found that the increased surface roughness in urban areas delayed the onset of the sea breeze by 1-2 hours and weakened its intensity by 20-40% compared to rural areas. This finding was supported by You et al. (2019), who investigated the impact of urbanization on the land-sea breeze circulation in the Pearl River Delta region of China. They found that the increased surface roughness and heat capacity of the urban areas weakened the land-sea temperature contrast and reduced

the strength of the onshore flow. This led to a reduction in coastal precipitation and an increase in inland convection, as the weakened sea breeze allowed the UHI circulation to dominate.

2.7 Synthesis and Research Opportunities

The literature review has highlighted significant progress in understanding how urban areas influence convective processes, from the initiation and evolution of storms to the resulting precipitation patterns (Shepherd, 2005; Han et al., 2014; Qian et al., 2022). However, a closer examination of the existing studies reveals a critical gap in our knowledge of the relationship between urban environments, surface heterogeneities, and the microphysical properties and vertical structure of convective cells. While previous observational studies have provided valuable insights into urban-induced rainfall anomalies (Shepherd et al., 2002; Niyogi et al., 2011), they have primarily focused on the overall impact of urban areas on precipitation patterns and storm frequency, but have not delved into the specific changes in storm characteristics as they evolve over time and interact with different surface environments. Moreover, although some studies have examined the vertical structure of convective cells using radar data (Bodine and Rasmussen, 2017; Feng et al., 2022; Marquis et al., 2023), there is still a need for in-depth analyses to compare the vertical characteristics of cells across the urban and surrounding rural regions. Investigating how the vertical distribution of microphysical properties and radar signatures change in response to different near-storm environments and surface heterogeneities is essential for understanding the mechanisms through which urban areas modulate convective cell evolution and precipitation formation.

This study aims to fill these critical gaps by harnessing dual-polarization radar measurements to explore the microphysical properties and vertical structure of convective cells in the Houston metropolitan area. The novel aspects of this study directly address the key research questions outlined in Chapter 1. By investigating the differences in the morphology and characteristics of convective cells across urban and suburban areas, and examining how these characteristics change as cells move through different regions, this study will provide crucial insights into the role of surface forcing and coastal interactions in shaping urban-influenced convection. Furthermore, by analyzing the near-storm environments throughout the lifecycle of convective cells, this study will shed light on how urban surface properties modulate instability, moisture, and lift, thereby influencing convective initiation and evolution.

The findings from this study are expected to significantly advance our understanding of urban-convection interactions, particularly in terms of microphysical processes and their vertical distributions. By providing a comprehensive assessment of how cell microphysics and structure evolve across different regions and in response to varying surface forcings, this study will contribute to improving the representation of these processes in numerical models and enhancing our ability to forecast urban-influenced convective storms. The next chapter will detail the specific data sources and methodology employed to address these research questions.

Chapter 3

Data & Methods

This chapter describes the datasets used in this study, along with the methods for identifying and tracking convective cells. It also explains the classification schemes for convective cell development stages, the analysis of vertical cell structure, and the delineation of the study area based on prevailing wind patterns.

3.1 Data Sources and Processing

This study utilized four distinct data types, as outlined in Table 3.1. Convective cell characteristics were derived from KHGX next-generation radar (NEXRAD) observations¹. Environmental metrics, indicative of conditions favorable or adverse to convective activity, were calculated from the Atmospheric Radiation Measurement (ARM) Interpolated Sonde (INTERPSONDE) data (Jensen et al., 2022)². Stage-IV precipitation dataset was also employed to investigate changes in storm intensity in the urban region relative to other regions. Additionally, land-use/land-cover classification (LULC) data were utilized to delineate the urban area from the entire study region to assess its impact on convective cell activity compared to other regions.

¹Radar data downloaded via ftp from <https://noaa-nexrad-level2.s3.amazonaws.com/index.html>

²https://adc.arm.gov/discovery/#/results/instrument_class_code::interpsonde

Table 3.1: Data sources and specifications

Data Type	Source	Resolution	Period
Radar	KHGX NEXRAD S-band	1° beamwidth, 250-m range resolution, 5 mins	June 2022 - August 2022
Soundings	INTERPSONDE	Varies with height, 1 min	June 2022 - August 2022
Stage-IV precipitation	NCAR UCAR EOL data	4 km, hourly	June 2022 - August 2022
LULC	USFS Landscape Change Monitoring System	30 m	June 2022 - August 2022

3.1.1 Radar

Level-II archived radar data were obtained from the National Centers for Environmental Information (NCEI) for the KHGX radar located in Houston, Texas (Fig. 3.1) (NOAA National Weather Service, 1991). The data covered the period from 1 June 2022 to 31 August 2022. The KHGX radar has a maximum range of 400 km and provides reflectivity and radial velocity measurements at temporal and range resolutions of about 5 mins and 250 m, respectively (Biswas and Chandrasekar, 2018). The radar data encompassed moment data such as reflectivity (Z), mean radial velocity, spectrum width, as well as dual-polarization data, including differential reflectivity (Z_{DR}), correlation coefficient (ρ_{hv}), and differential phase (K_{DP}). The data were reported at intervals of 4.5, 5, 6, or 10 mins, depending on the volume coverage pattern employed (Priftis et al., 2021).

In this study, we mapped the radar data to a three-dimensional Cartesian grid with horizontal and vertical resolutions of 500 m and a temporal frequency of approximately 5 mins using the Python ARM Radar Toolkit (Py-ART) (Helmus and Collis, 2016).

The Barnes2 objective analysis scheme (Pauley and Wu, 1990) was employed as the interpolation method, which accounts for the non-uniform distribution of radar data and the inherent spatiotemporal correlation of meteorological variables. To ensure the quality of the gridded data, strict quality control measures were implemented. Specifically, radar gates with ρ_{hv} below 0.9 were excluded from the analysis, as low ρ_{hv} are often indicative of non-meteorological echoes (Kumjian, 2013a). While meteorological echoes, such as large, wet hail, can be below 0.9, large hail was not encountered during the study period. This filtering step helped mitigate the influence of boundary layer echoes, ensuring that the remaining data primarily constitutes radar measurements pertaining to the convective cells of interest.



Figure 3.1: Image of the NEXRAD S-band radar³

Additionally, K_{DP} was derived from the dual-polarization radar measurements using Py-ART. K_{DP} is the range derivative of the differential propagation phase shift, and it is a valuable quantitative estimator of rainfall rate. Unlike Z , which is affected by the presence of large drops, hail, and partially melted hydrometeors, K_{DP} is less sensitive to these factors. K_{DP} is directly related to the mass of liquid water in the

³<https://www.roc.noaa.gov/WSR88D/Engineering/NEXRADTechInfo.aspx>

radar resolution volume, making it a more robust estimator of rainfall rate, especially in heavy precipitation events (Bringi and Chandrasekar, 2001).

3.1.2 Soundings

The sounding measurements were obtained from the ARM INTERPSONDE value-added product (VAP). The INTERPSONDE VAP transforms discrete radiosonde measurements into continuous daily files, presenting atmospheric thermodynamic parameters on a fixed time-height grid. It utilizes sounding data and precipitable water vapor estimates from microwave radiometer measurements to linearly interpolate atmospheric profiles at a temporal resolution of 1 min, spanning 332 vertical levels from the surface up to approximately 40 km altitude (Fairless et al., 2021). The INTERPSONDE VAP addresses the limitation of traditional radiosonde data, which is typically available only every 12 h. By interpolating the atmospheric state variables linearly in time for each height level between sounding times, the VAP offers a more comprehensive representation of atmospheric conditions, providing estimates at higher temporal resolutions. It faithfully preserves the original radiosonde measurements at the exact sounding times, ensuring the integrity of the observational data (Fairless et al., 2021). In this study, the INTERPSONDE VAP data for the Houston site were utilized to calculate the environmental conditions near the convective cells (Jensen et al., 2022). The VAP’s continuous time-height representation of atmospheric parameters enables us to retrieve relevant atmospheric conditions and stability indices at a higher temporal resolution than standard radiosonde soundings. This high temporal resolution captures the rapidly evolving atmospheric state surrounding the cells, enabling us to analyze the environmental factors influencing their evolution.

3.1.3 Stage-IV Precipitation

The Stage-IV product (Lin and Mitchell, 2005) is a multi-sensor analysis that integrates radar observations from approximately 140 radars and rain gauge measurements from around 5,500 stations across the contiguous United States (CONUS). The analysis process begins with regional precipitation analyses produced by the 12 River Forecast Centers, each employing quality control procedures and bias correction techniques to optimally merge radar and gauge data within their respective domains. These regional products are then mosaicked into a national precipitation analysis by the National Centers for Environmental Prediction. For this study, the Stage-IV precipitation analysis product covering the Houston area for the period of June-August 2022 was utilized (Du, 2011). The data provides high-resolution, area-averaged precipitation estimates on a 4 km x 4 km grid over the CONUS at an hourly temporal resolution. A key advantage of the Stage IV product is its ability to better characterize convective precipitation systems compared to rain gauge observations alone (Nelson et al., 2016). Unlike sparsely distributed gauge networks, which may undersample localized convective events, the Stage IV product leverages the spatial coverage provided by radar observations to capture the structure and evolution of convective precipitation more effectively.

3.1.4 Land Cover

Land cover information for the study area was obtained from the United States Forest Service (USFS) Landscape Change Monitoring System (LCMS) version 2021.7 (Housman et al., 2021). The LCMS is a comprehensive remote sensing-based system that provides continuous monitoring of vegetation cover change, land cover, and land use across the CONUS and southeastern Alaska. It utilizes Landsat and Sentinel-2 data in the Landsat-based detection of Trends in Disturbance and Recovery (LandTrendr)

and Continuous Change Detection and Classification (CCDC) temporal segmentation algorithms to map changes over time. The LCMS employs a hierarchical classification scheme to categorize land cover and land use into multiple thematic classes, including agriculture, developed, forest, non-forest wetland, other, and rangeland or pasture. The classification process involves supervised random forest models trained on calibration data collected through the TimeSync attribution tool, which allows for the attribution of change processes, land cover, and land use at sample locations using multi-temporal Landsat imagery and ancillary data (Cohen et al., 2010).

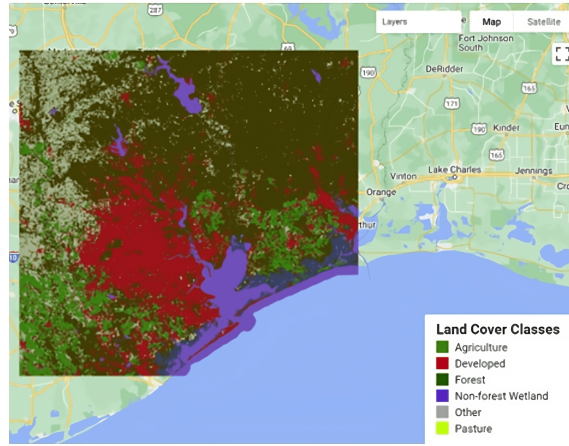


Figure 3.2: Land cover classification

The LCMS data products are generated through a robust processing chain, which includes image preprocessing, cloud and cloud shadow masking, multi-temporal compositing, temporal segmentation using LandTrendr and CCDC, predictor variable selection, and supervised classification. The resulting maps offer a spatial resolution of 30 m and are updated annually from 1985 to the most recent season. Quality assurance procedures, such as model validation and cross-validation, are implemented to ensure the accuracy and consistency of the products. For this study, the “Developed” class from the LCMS land use data for the year 2022 was utilized (Fig. 3.2). This class

represents areas with man-made structures and mixed vegetation/structures altered by human activities. We extracted the shapefile of the class within the Google Earth Engine platform and used it to define the urban area within the study region.

3.2 Description of the Study Area

Houston is the largest city in the southern United States (Zhang et al., 2018). Its location provides an excellent opportunity to examine the variability of atmospheric parameters across convective cell lifetimes. The city is situated in a humid subtropical climate zone, where onshore flow and sea-breeze convection interact, resulting in diverse atmospheric conditions (Shepherd et al., 2010). Updrafts in Houston are mainly driven by the convergence of warm air with the Gulf’s sea breeze, creating favorable conditions for thunderstorm development (Fridlind et al., 2019). The region’s climate is influenced by moisture from the Gulf of Mexico, leading to hot, humid summers and cool, wet winters (Statkewicz et al., 2021). Houston’s prevailing wind is from the south-southeast direction, with a mean surface speed of 7.4 mph, further emphasizing the dominance of the Gulf’s influence on the area’s atmospheric dynamics⁴. The city’s topography is relatively flat, with elevations ranging from sea level to around 125 feet (38 m) above sea level (Yu et al., 2009). This flat terrain allows for the unimpeded flow of moist air from the Gulf, contributing to the development of convective storms.

⁴<https://www.visithoustontexas.com/aroundhouston/weather/>

3.3 Methodology

3.3.1 Definition of the Coordinate System

To investigate potential urban impacts on precipitation patterns, this study employed a modified coordinate system framework (Fig. 3.3) inspired by Huff and Changnon (1972) and Shepherd et al. (2002). Instead of relying solely on climatological averages, we utilized the prevailing lower-tropospheric wind flow to delineate the various regions. This approach enables us to capture potential urban precipitation signatures while accounting for the influence of the prevailing wind flow on the spatial distribution of cells in the hypothesized impact regions.

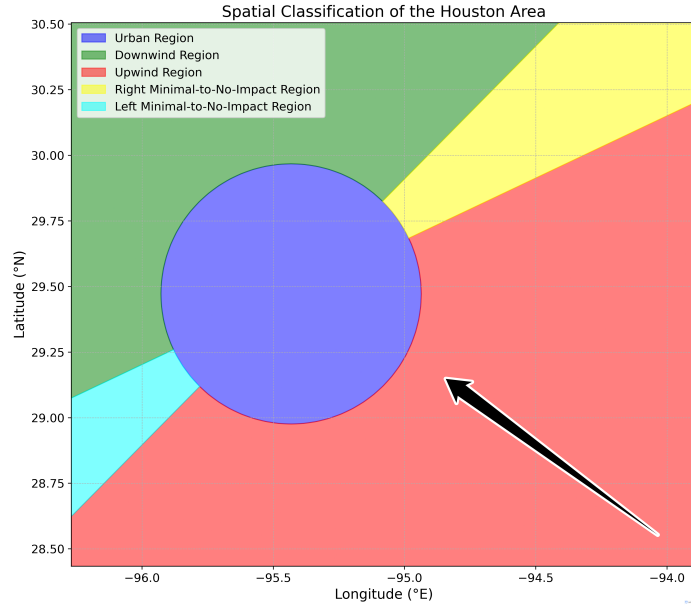


Figure 3.3: Regional classification of the Houston metropolitan area and its surroundings. The red circle delineates the upwind region, while the red wedge represents the upwind sector. The green wedge indicates the downwind sector. The blue and yellow sectors serve as buffer zones. The black arrow denotes the direction of the prevailing wind flow across the region.

The orientation of this coordinate system is determined by the mean 700 hPa wind vector, which represents the average steering flow for convective systems during the study period. This steering level was chosen based on previous studies that have identified 700 hPa as a representative level for the mean wind flow influencing convective storm propagation (Carbone et al., 2002; Feng et al., 2018; Hagemeyer and Schmocker, 1991). The mean 700 hPa wind direction for June-August 2022 was calculated using ERA5 reanalysis dataset (Copernicus Climate Change Service, 2018). The coordinate system is centered on the urban area, with the following key components:

1. A central urban region covering the developed region of the Houston area. Here, precipitation enhancement due to UHI effects is anticipated.
2. A maximum impact zone located 25-75 km downwind of the urban center, spanning a 160° sector and aligned with the mean wind direction. This sector width accounts for the potential variability in the steering flow during individual events. This region is hypothesized to experience the most pronounced UHI-related precipitation impacts due to the advection of urban-modified air masses by the prevailing flow.
3. An upwind zone located 25-75 km upwind of the urban center, serving as a reference area and is unaffected by urban influences. This region is closer to the coast where the sea breeze will often initiate thunderstorms.
4. Lateral minimal-to-no-impact zones approximately 50 km orthogonal to the mean wind vector. These zones are expected to exhibit minimal or negligible urban impacts due to the lack of direct advection from the urban area.

The analysis focused primarily on the upwind, urban, and downwind regions, as they are of key interest for assessing the potential urban precipitation effects relative to

the upwind control region. We computed the average 700-hPa wind direction as 125.37°. Since this flow is south-southwesterly, we hypothesize that sea breeze circulation may interact with the prevailing synoptic flow and influence the spatial distribution and characteristics of convective cells, particularly in the urban and downwind zones.

To account for the varying spatial extents of the different regions, we calculated their respective areas (Table 3.2) and normalized the convective cell counts and precipitation amounts accordingly. This normalization step ensures a fair comparison across regions by considering their different sizes, enabling a more robust assessment of potential urban impacts on the frequency, intensity, and spatial distribution of the cells.

Table 3.2: Land area of the regions

Region	Area (km ²)
Upwind	25,178
Urban	8,297
Downwind	13,037
Left minimal-to-no-impact	4,910
Right minimal-to-impact	1,576

3.3.2 Cell Identification and their Near-Storm Environments

To track convective activity within the KHGX NEXRAD data, we utilize an object-based cell-tracking algorithm called PyFLEXTRKR (Feng et al., 2023). This algorithm differs from most existing packages as it is designed to be versatile, capable of tracking various types of storms across multiple scales. It explicitly handles situations where features merge or split, and it offers functionalities for visualization, post-processing, and analysis of the tracked features. Furthermore, PyFLEXTRKR incorporates scalable

parallelization options and performance optimizations, making it suitable for processing large datasets.

First, we converted the native polar coordinates of the 3D radar reflectivity information in the KHGX radar data to Cartesian coordinates while maintaining 500-m grid spacings using PyART software (Helmus and Collis, 2016). The grid spacing was selected to balance computational efficiency and preserve detail, as a finer grid spacing would require more computational and storage resources.

We then identified convective cells from the radar data using composite reflectivity (Fig. 3.4), which is the highest radar reflectivity in a column. This product is crucial because it enables us to detect the cells during their initial developmental stages as precipitation echoes are often detected first above the base elevation. Specifically, we utilized an adapted version of the horizontal radar reflectivity texture algorithm (Steiner et al., 1995), which focuses on the difference between the reflectivity at a specific grid point and the average reflectivity in its immediate vicinity. Following Feng et al. (2022), we classified a grid point as convective based on specific criteria:

1. Composite reflectivity > 50 dBZ, or
2. Difference between composite reflectivity and background composite reflectivity

(Z_{bkg}) is:

$$\Delta Z = \begin{cases} 10 \cos\left(\frac{\pi Z_{\text{bkg}}}{2 \times 60}\right), & \text{if } 0 \leq Z_{\text{bkg}} \leq 60 \text{ dBZ} \\ 0, & \text{if } Z_{\text{bkg}} > 60 \text{ dBZ} \end{cases} \quad (3.1)$$

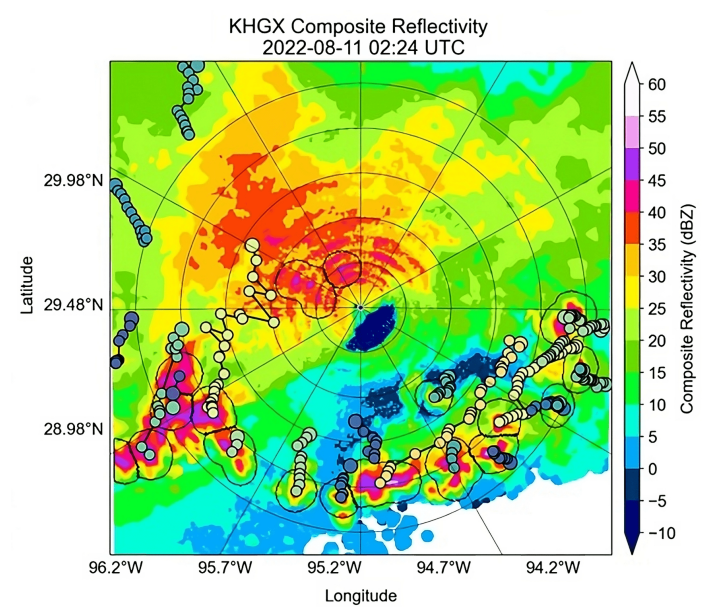


Figure 3.4: An example of the tracked convective cells from composite reflectivity on 11 April 2022

To effectively distinguish individual convective cells within “convective complexes” and prevent misidentification of convection in stratiform radar brightbands, a convective core covering a minimum of 16 grid points (or maximum core area of 4 km^2) is defined and then expanded using a Z_{bkg} -dependent radius step function (Feng et al., 2022). The convective cores are then expanded using a convective radius increment of 0.5 km, starting from a radius of 1 km up to a maximum radius of 5 km. Grid points near the convective cores with reflectivity values exceeding the truncated convective threshold of 55 dBZ are also classified as convective. Several aspects of the three-dimensional cell properties are then saved throughout the cell lifetime, including cell area, echo-top height, and maximum reflectivity. Table 3.3 presents some key parameters of the PyFLEXTRKR software used in this study.

Table 3.3: Key PyFLEXTRKR parameters and values

Parameter	Name	Value	Description
absConvThres	Absolute convective threshold (dBZ)	50	Minimum composite reflectivity value for a grid point to be classified as convective
minZdiff	Minimum reflectivity difference (dBZ)	10	Minimum difference between a grid point's composite reflectivity and the background reflectivity for it to be classified as convective
truncZconvThres	Truncated convective threshold (dBZ)	55	Reflectivity threshold for grid points near convective cores to be classified as convective
mindBZuse	Minimum dBZ for convective classification	25	Minimum reflectivity value for a grid point to be considered for convective classification
conv_rad_increment	Radius increment (km)	0.5	Increment for expanding convective cores
bkg_refl_increment	Background reflectivity increment (dBZ)	5	Increment for calculating background reflectivity
maxConvRadius	Maximum convective radius (km)	5	Maximum allowed radius for convective cores
bkgndRadius	Background radius (km)	11	Radius for calculating background reflectivity intensity
min_corearea	Minimum convective core area (km ²)	4	Minimum area required for a group of grid points to be considered a convective core

To characterize the near-storm environment of the cells, we calculated several sounding metrics relevant for convective activity in addition to the base atmospheric quantities using the INTERPSONDE data. The sounding metrics were calculated at various standard heights and over vertical layers relevant to different cloud processes, and they include convective available potential energy (CAPE), convective inhibition (CIN), bulk shear, and storm relative helicity. Specifically, we focused on the mixed layer CAPE and CIN within the lowest 50 hPa, the most unstable layer CAPE and CIN within the lowest 50 hPa, and the storm relative helicity and bulk shear over the 0-1 km, 0-3 km, and 0-6 km layers. The mixed layer within the lowest 50 hPa provides information about the initial instability and moisture content that can fuel convective initiation. These layers capture crucial information about the low-level moisture, instability, and wind shear, which are key drivers of convective initiation, organization, and evolution. The most unstable layer within the lowest 50 hPa captures the level of maximum buoyancy and instability, which is a key factor in determining the potential strength and updraft velocities of thunderstorms. Storm relative helicity (SRH) and bulk shear over the 0-1 km, 0-3 km, and 0-6 km layers are important for assessing the potential for organized rotation and thunderstorm development.

In addition to these layers, we also calculated wind speeds, dewpoint temperature, and relative humidity at the surface, 500 hPa, and 850 hPa levels. The surface level represents the conditions near the ground, which influence the initial development of convective cells. The 500 hPa level corresponds to the mid-tropospheric region, where the strength and organization of convective storms are often determined. The 850-hPa level is situated within the lower troposphere, capturing crucial information about moisture and instability, which are key drivers of convective initiation and evolution.

Subsequently, we composited the calculated sounding data to the base times of the cells identified by the PyFLEXTRKR algorithm. To ensure proper compositing, we

first grouped the cells by their base times. Since the base times of the cells contain milliseconds, we rounded up each base time to the nearest 30 seconds. By grouping the cells and rounding their base times, we create a consistent temporal framework for the compositing process. Next, we matched each group of convective cells with the nearest available sounding data in time. This approach ensures that the sounding metrics are temporally and spatially consistent with the observed convective activity, providing a reliable representation of the near-storm environment for each group of cells.

3.3.3 Classification of Cell Development Stages

The evolution of convective storm cells is characterized by distinct phases of development, each exhibiting specific characteristics related to the cell's size, intensity, and vertical structure. To facilitate the characterization of convective cell properties during different lifecycle stages, a consistent classification scheme based on relevant meteorological parameters was established in this study:

1. Developing stage: A cell is classified as “developing” if there is a consistent and significant increase in the cell area, reflectivity, and echo-top height (in this case, the 20-dBZ echo top height) over multiple consecutive radar scans. The magnitude of these changes should be notably larger than the typical scan-to-scan variability observed during the mature stage. This indicates that the cell is growing steadily in both size and intensity.
2. Mature stage: A cell transitions from the “developing” stage to the "mature" stage when the rate of change in its area, reflectivity, and echo-top height decreases significantly and remains relatively stable over several radar scans. While small fluctuations in these parameters may still occur from scan to scan, the overall trend should be much flatter compared to the developing stage. The cell's area,

reflectivity, and echo-top height are expected to be at or near their peak values during this stage, indicating that the cell has reached its maximum development.

3. Dissipating stage: A cell is classified as "dissipating" when it exhibits a consistent and significant decrease in area, reflectivity, and echo-top height over multiple consecutive radar scans. The magnitude of these changes should be comparable to or larger than the changes observed during the developing stage, but in the opposite direction. This decline indicates that the cell is steadily weakening and losing its intensity and size.

3.3.4 Vertical Distribution of Convective Cell Properties

To examine the vertical distribution of the cell properties, this study employs contoured frequency by altitude diagrams (CFADs) (Yuter and Houze, 1995). CFADs are a powerful tool for visualizing and analyzing the statistical distribution of radar-derived parameters as a function of altitude. They essentially represent a series of histograms plotted at different heights, revealing the distribution of radar variables at each altitude level rather than a single distribution or statistic. This approach is particularly useful when dealing with complex radar echo patterns, as it enables the examination of ensemble properties across multiple convective cells, rather than focusing on individual entities.

In this study, we construct CFADs to investigate the vertical structure and evolution of convective cells at different stages of their development. First, we sample the gridded radar data associated with the tracked convective cells along the vertical column for each cell, extracting Z , ρ_{hv} , Z_{DR} , and K_{DP} . The gridding process is necessary because it addresses the spatial bias inherent in the raw polar radar data, where the radar beam widens with increasing distance from the radar, resulting in larger resolution volumes

and a bias towards values near the radar where a storm would have more resolution volumes compared to a farther range storm. By interpolating the data onto a Cartesian coordinate system with uniform spacing, the gridding process ensures that each data point contributes equally to the histogram, regardless of its distance from the radar.

Subsequently, we binned the sampled data into predefined intervals (Table 3.4) and calculated the frequency of occurrence for each bin at each height level. The bin intervals are chosen to provide sufficient resolution for capturing the variability in each parameter while maintaining computational efficiency. This process is repeated for all convective cells within the analysis period, resulting in a two-dimensional frequency distribution of the parameter of interest as a function of height. To facilitate interpretation and comparison between different datasets or regions, we normalized the CFADs by the maximum count across all heights to remove the effect of the bulk amount of radar echo in a given subset of the data, as well as to bring the maximum in the profile to the same magnitude for each computed CFAD while not changing the shape of the distribution (Hence and Houze, 2011).

Finally, the normalized frequency distribution is plotted as a CFAD, with the x-axis representing the parameter value (e.g., reflectivity in dBZ) and the y-axis representing altitude. The contours on the CFAD indicate the relative frequency of occurrence for each parameter value at each height level.

Table 3.4: CFAD bin intervals

Parameter	Bin interval
Z	5 dBZ
Z_{DR}	0.5 dB
ρ_{hv}	0.0025
K_{DP}	0.1 deg/km

Chapter 4

Spatiotemporal Characteristics and Environmental Factors of Convective Cells

4.1 Spatiotemporal Characteristics of Convective Cells

A total of 22,372 cell tracks (141,453 individual cells) were analyzed across the Houston, Texas region: 10,906 tracks (73,281 cells) in the upwind region, 7,446 tracks (45,051 cells) in the urban region, and 4,020 tracks (23,121 cells) in the downwind region (Table 4.1). In this context, a ‘cell’ refers to any individual convective cell detected at a single time step, while a ‘track’ represents the temporal evolution of a cell or group of cells linked together over time. This distinction allows us to capture both the instantaneous presence of cells and their lifecycle throughout the study period.

Table 4.1: Distribution of cell tracks across different regions

Tracks	All regions	Upwind region	Urban region	Downwind region
Total	22,372	10,906	7,446	4,020
Non-merge	18,317	9,066	6,016	3,235
Merge	4,055	1,840	1,430	785
Non-split	18,454	9,141	5,970	3,343
Split	3,918	1,765	1,476	677

The study period spanned the summer months of June–August 2022, with the average and total rainfall amounts increasing month by month (Table 4.2). Out of the total 22,372 cell tracks, 17,793 (81.9%) were non-merging cell tracks and 4,055 (18.1%) were merging cell tracks. Similarly, 18,454 (82.5%) were non-splitting tracks while 3,918 (17.5%) were splitting tracks. The proportions of merging and splitting tracks were relatively consistent across the upwind, urban, and downwind regions, with the urban region having slightly higher percentages of both merging (19.2%) and splitting (19.8%) cells compared to other regions.

Table 4.2: Distribution of rainfall across the summer months

Month	Average rain amount (all values; mm)	Average rain amount (non-zero values; mm)	Total rain amount (mm)
June	0.079	2.473	112,463.688
July	0.131	2.436	137,087.438
August	0.280	2.282	410,984.375

The frequency analysis of the cells was calculated using the cell tracks. In this case, if a cell splits, the resulting cells are considered as separate and independent tracks, and if cells merge, the merged cell is considered a new track. This methodology

allows for a more accurate representation of the lifecycle and duration of the cells, as it accounts for cell splitting and merging events. By focusing on the tracks, we can better investigate the temporal evolution of the cells and compare their characteristics across different regions without the potential limitations of double-counting splitting cells or underestimating cell duration during mergers.

4.1.1 Empirical Distribution of Cells in the Upwind Region

The spatial distribution of the cell track locations in the Houston area reveals a relatively high number of cells initiated in the upwind region near the coast (Fig. 4.1(a)). This finding is consistent with previous studies that have demonstrated the strong influence of sea breezes on convection initiation in coastal environments (Fovell, 2005; Tuftedal et al., 2023). The effect of sea breezes on convection initiation in the upwind region is particularly pronounced overnight to early morning hours (Fig. 4.2(c)-(e), 4.9(a)). This phenomenon is likely due to the interaction between the nighttime land breeze moving over the Gulf and the incoming sea breeze, which creates a zone of enhanced convergence that favors the development of cells even in the absence of peak daytime heating (Allouche et al., 2023). The dominance of early morning initiation in the upwind region may significantly influence the behavior of the resulting cells, causing them to exhibit different structure and lifecycle compared to those initiated further inland.

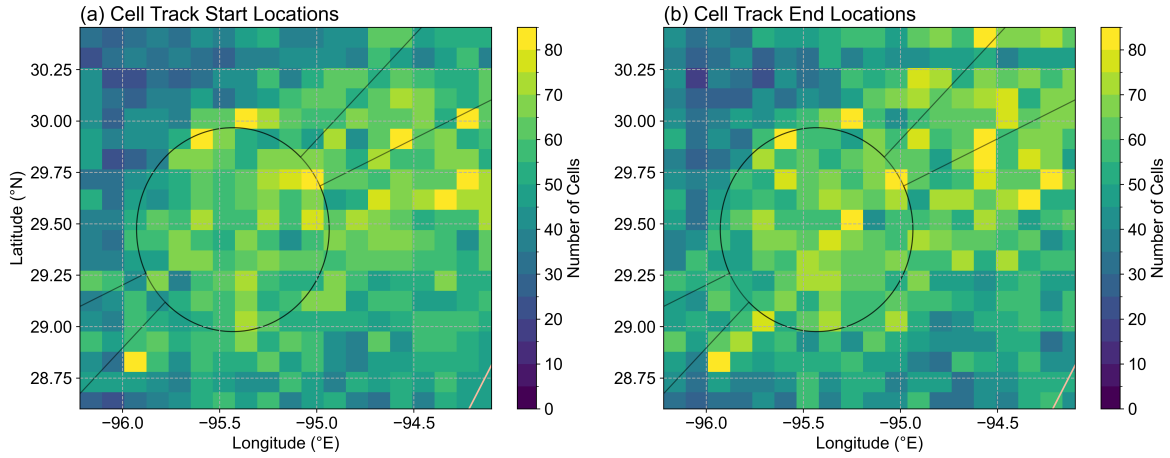


Figure 4.1: Spatial distribution of convective cell track locations in the Houston area. (a) Cell track start locations, showing the cell initiation points. (b) Cell track end locations, showing where the cells dissipate. The urban region is indicated by the circle, with the upwind and downwind regions positioned relative to prevailing south-southeasterly wind direction. The color scale represents the number of cells.

The upwind region also experienced high total precipitation in June–August, 2022, with a median value of around 130 mm and a mean slightly above 130 mm. While the similarity between the mean and median precipitation values suggests a relatively symmetric distribution, the moderate spread and interquartile range indicate varying rainfall intensities. Most rainfall events in this region likely result in moderate amounts, as suggested by the interquartile range. However, the maximum values of up to 225 mm over the summer of 2022 point towards occasional heavy rainfall events, potentially driven by the regular influence of sea breezes.

Meanwhile, the moderate total precipitation per cell (20 mm), cell density (8 cells per 10 km² land area), average maximum 20-dBZ echo top heights (8 km), and average maximum reflectivity values (36 dBZ) (shown in Fig. 4.4) indicate steady, moderate convective activity driven by sea breeze circulations. Moreover, the smaller average

maximum cell areas observed in this region (42.5 km^2) compared to the urban and downwind regions suggest that cells in this area are comparatively less intense and smaller in scale (Fig. 4.4(c)). The lifetime distribution of cells in this region (Fig. 4.5) provides additional evidence for the dominant role of sea breezes in convective activity. The high frequency of cells and the rapid decrease in cell frequency around a lifetime of 5 hours suggest that most cells in this region last for less than 5 hours, consistent with the localized and transient nature of sea breeze circulations (Tuftedal et al., 2023).

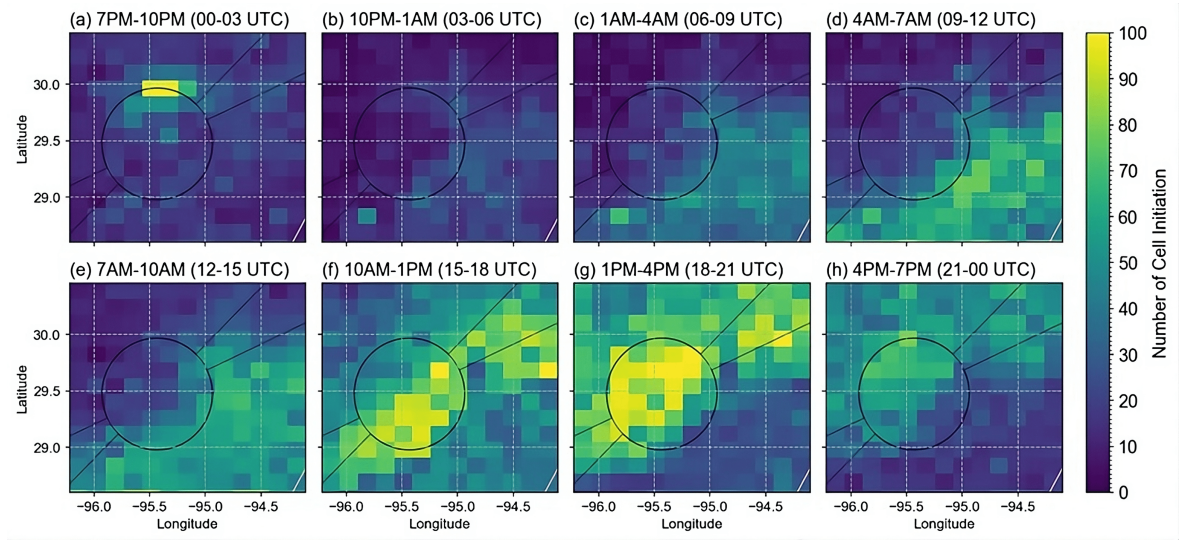


Figure 4.2: Diurnal variation of convective cell initiation in the Houston area

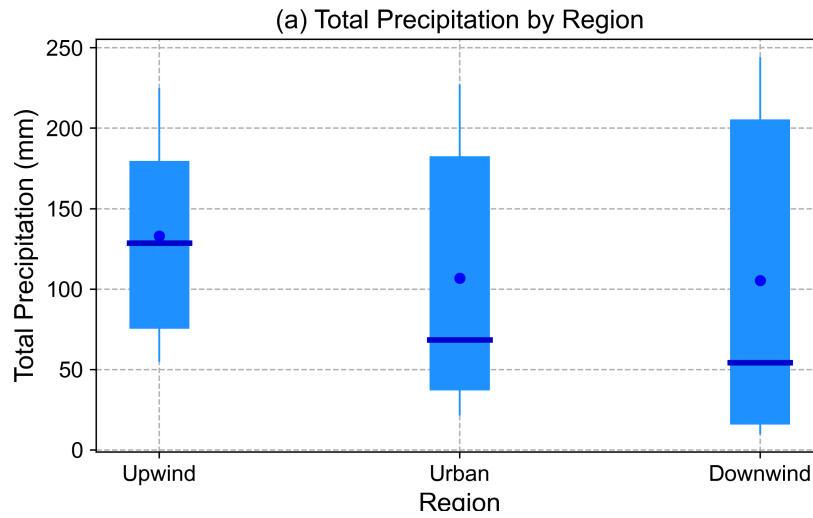


Figure 4.3: Average total precipitation by region. Plots show the distribution of precipitation (mm) for upwind, urban, and downwind regions. The thick line represents the median, and the dot indicates the mean. The whiskers show the range, and they extend from the 10th percentile to the 90th percentile to exclude potential outliers beyond these limits.

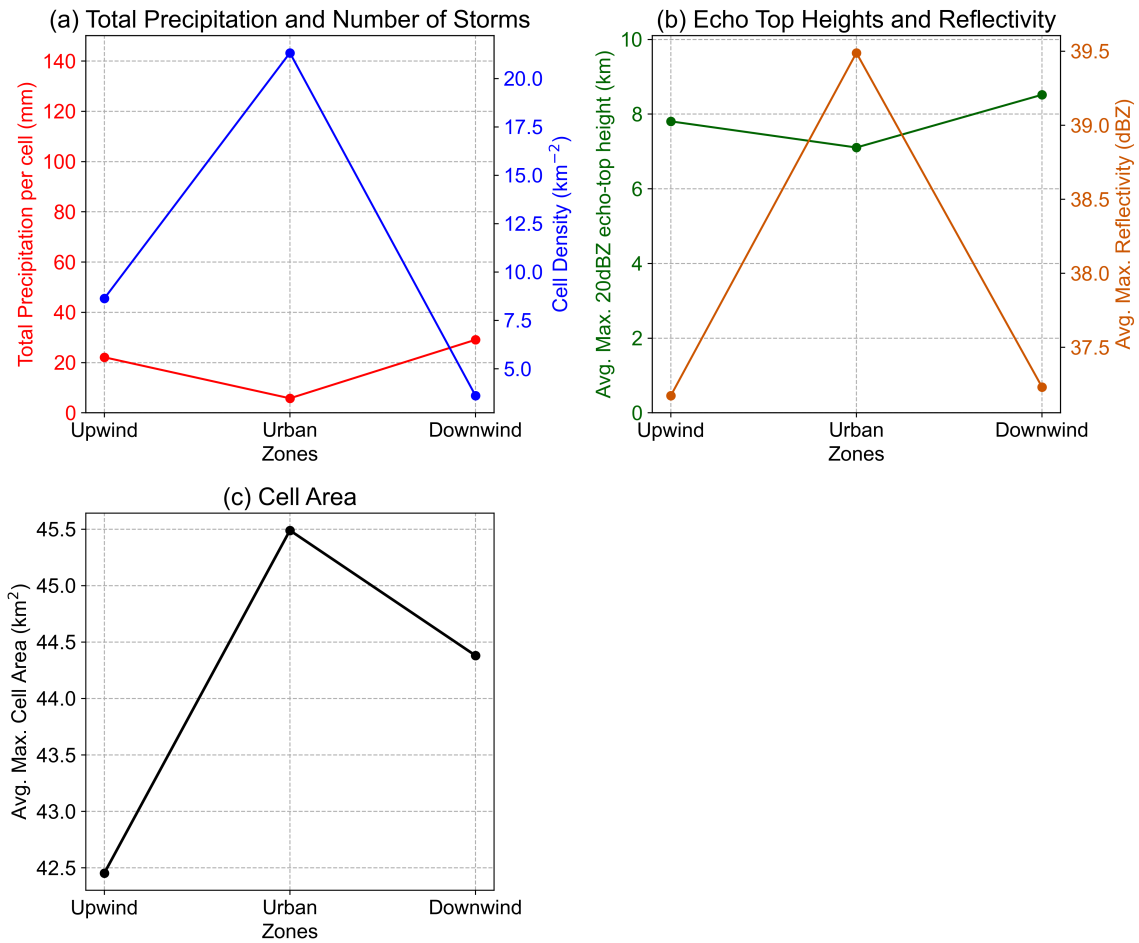


Figure 4.4: Comparison of convective cell characteristics across regions. (a) Average total precipitation per cell (red) and cell density (blue); (b) Average maximum 20-dBZ echo top heights (green) and reflectivity (purple); (c) Average maximum cell area for each zone. Cell density is the number of cells per 10 km^2 land area. All “average maximum” metrics represent the mean of peak values observed for each storm cell during its lifetime.

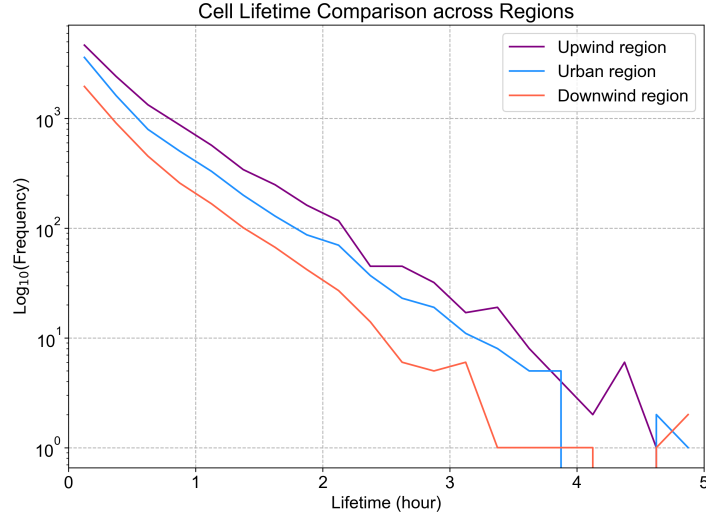


Figure 4.5: Cell lifetime comparison across regions in the Houston area

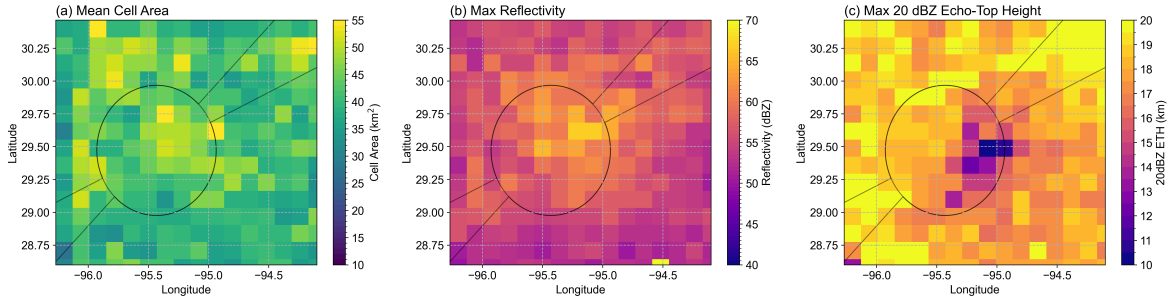


Figure 4.6: Spatial distribution of convective cell characteristics in the Houston area. (a) Mean cell area, indicating the average size of convective cells in each region. (b) Maximum reflectivity, showing the intensity of convective cells. (c) Maximum 20-dBZ echo-top height, representing the vertical extent of the cells.

4.1.2 Empirical Distribution of Cells in the Urban Region

Compared to the downwind region, the urban region has a relatively higher number of initiated cells (Fig. 4.1(a)). This is attributed to the UHI effect, which increases surface

temperatures and enhances convection initiation, particularly during the afternoon to evening hours when urban heating is maximized (Fig. 4.2(f-g)). The built environment and anthropogenic heat sources in the urban area create localized hotspots that further contribute to the development of convective cells (Haberlie et al., 2015). Shepherd et al. (2010) found similar results for Houston and maintained that the UHI destabilized the boundary layer, providing the required buoyancy to support convection initiation. Interestingly, the urban region also shows a higher concentration of cell dissipation events than both the upwind and downwind regions (Fig. 4.1(b)), suggesting that some cells may dissipate within or near the urban environment. This could be due to various factors, such as localized downdrafts or interactions with urban structures that influence airflow and stability.

One notable observation is high frequency of cells with lifetimes less than 4 hours, which is 1 hour shorter than those in the upwind region (Fig. 4.5). This suggests that while the UHI effect may be enhancing convection initiation in this region, it is not necessarily sustaining it. Instead, it appears to be frequently disrupting the development of existing convective cells, leading to their shorter lifetimes. The disruption could occur through shifting areas of low-level convergence and lift as the UHI circulation evolves, removing the forcing that was sustaining the cells. Another plausible explanation could be enhanced turbulence owing to increased surface roughness, which can disrupt the organized updrafts necessary for sustaining the cells. The shorter lifetimes may also contribute to the higher variability in precipitation patterns observed in this area (Fig. 4.3). The median precipitation is around 65 mm, with the mean slightly above 100 mm, suggesting the occurrence of both periods or areas of less intense precipitation and substantial rainfall events. Shorter-lived cells are less likely to produce prolonged precipitation events, leading to a higher frequency of lighter rainfall events. However, when conditions are favorable for cell mergers or interactions, these shorter-lived cells

can intensify and produce heavier rainfall amounts, contributing to the observed wide range of precipitation values in the urban region.

The cell characteristics in this region, such as low total precipitation per cell (5 mm), high cell density (20 cells per 10 km² land area), low average maximum 20-dBZ echo top height (7 km), high average maximum reflectivity (39.5 dBZ), and large average maximum cell area (45.5 km²) (Fig. 4.4), provide further insights into the nature of urban-influenced convection. The low precipitation per cell, despite the high cell density, suggests that many cells are initiated in this region but may be less intense individually. The lowest average maximum echo top height in this region also indicates that while the cells are intense, they may not grow as tall, possibly due to limitations in vertical development within the urban environment. However, the highest average reflectivity values and largest average cell areas in this region compared to other regions suggest that the UHI effect and cell merging contributes to the intensification and horizontal growth of these cells (Fig. 4.6) (Tuftedal et al., 2023).

When interpreting the cell characteristics in the urban region, it is important to consider the potential influence of the radar scanning mechanism on the observed patterns. The analysis above reveals a reduced frequency in the average 20-dBZ echo top height near the right flank of the urban area (Fig. 4.6(c)). This could potentially be an artifact or bias caused by the radar's inability to sample the full depth of the storm near the radar site. The radar may be under-sampling the true echo top heights in that area and only detecting echoes up to the limit of its highest elevation scan, making it appear that the urban-induced cells are lower in height than they truly are. However, the unique nature of this pattern suggests that other factors may also be contributing to this observation in that specific area. The fact that the reduced echo top heights are concentrated at the right flank of the urban area and not consistently observed in other parts of the region equidistant from the radar indicates that local

factors, such as atmospheric conditions, may also be influencing the cell characteristics in that area. Despite the potential influence of the radar scanning mechanism and the fact that the absolute cell areas may be overestimated at farther ranges due to beam broadening, the relative differences between regions remains valid. When compared to the upwind and downwind regions, the convection initiation, cell, and precipitation characteristics observed in the urban region provide strong evidence for the significant impact of the UHI effect on convective processes. The consistency of these findings with previous studies on urban-induced convection further supports the reliability of the results (Niyogi et al., 2011; Shepherd et al., 2010; Tuftedal et al., 2023).

4.1.3 Empirical Distribution of Cells in the Downwind Region

The downwind region is influenced by the advection of cells from the urban area and their subsequent evolution. The results show a lower number of cell dissipation events in this region compared to the urban region, suggesting that some cells may terminate within the urban area before reaching the downwind region, while others may merge into new cells. This merging process can lead to fewer but potentially larger cells (Fig. 4.6(a)). The merging of cells is a key process in the downwind region, as evidenced by the larger cell areas observed in this region compared to the upwind region (Fig. 4.6(a)). This indicates the growth and intensification of the cells, with the process driven by interactions between the advected cells and local atmospheric conditions, such as enhanced instability. Meanwhile, this region exhibits larger average cell areas compared to the upwind region (Fig. 4.6(a)); however, the average cell areas are generally smaller than those in the urban region, suggesting that while cells continue to grow and merge in the downwind region, they may not attain the same sizes as those in the urban area. This could be due to the gradual dissipation of the UHI effect and the changes in surface characteristics. As the cells move away from the urban

area, the UHI effect gradually diminishes, as it is typically most intense within the city boundaries and decreases in intensity with increasing distance from the urban core. Consequently, the additional buoyancy and instability provided by the UHI effect may be reduced in the downwind region, limiting the potential for further cell growth and intensification.

Interestingly, the downwind region exhibits some of the highest echo-top heights (Fig. 4.6(c)), suggesting that cells in this region can achieve significant vertical development (Note: The 20-dBZ echo top heights in the downwind region may be slightly underestimated due to beam overshooting effects at greater distances from the radar. This is discussed in more detail in Section 4.5.1). This vertical growth is likely supported by favorable atmospheric conditions, such as continued instability. The lifetime distribution of convective cells in the downwind region (Fig. 4.5) reveals longer cell lifetimes in this region compared to the urban region but with fewer cells overall. This suggests that while cell initiation may be less frequent in the downwind region, the environmental conditions are more favorable for sustaining the development and persistence of those cells that do form, allowing them to maintain their longevity. The precipitation characteristics in this region highlight the continued development of the cells (Fig. 4.3). The results reveal a considerable difference between the mean and median precipitation values, indicating a wide range of rainfall amounts. This variability can be attributed to the occurrence of both moderate rainfall events associated with the sustained convective cells and extreme rainfall events resulting from cell mergers, interactions, or particularly intense and long-lived cells.

Moreover, the downwind region has the highest total precipitation per cell (30 mm) among the three regions (Fig. 4.4), indicating the greater efficiency of the cells in producing rainfall. Despite having a lower density (2 cells per 10 km² land area), the cells in this region tend to be more organized and intense, capable of producing substantial

rainfall amounts. The high average maximum 20-dBZ echo top height (9 km) observed in this region (Fig. 4.4) are also indicative of strong updrafts and deep convection, which are necessary for the formation of severe weather and heavy precipitation. This observation aligns with the previous explanations about the continued vertical growth and intensification of convective cells in the downwind region. Although slightly lower than the urban region, the average maximum reflectivity (36.6 dBZ) in this region remains high, indicating the persistence of intense convective activity in this area. The slightly smaller average maximum cell area (44.5 km²) compared to the urban region (Fig. 4.6(c)) suggest that while cells may continue to merge and grow, the growth is not as extensive as in the urban area, potentially due to the lower cell density.

4.1.4 Analysis of Common Cell Tracks and Changes in Cell Properties

In Sections 4.1.1-4.1.3, we analyzed the characteristics of cells within the upwind, urban, and downwind regions of Houston. While this approach provided valuable insights into the general cell properties in each region, it did not explicitly examine the evolution of specific cell tracks as they moved across the regions. To complement these findings and investigate the potential impact of surface heterogeneities on the evolution of individual cells, we now focus on the analysis of cell tracks and changes in their properties as they traverse different regions. Specifically, we identified common cell tracks across these regions and examined changes in cell area, average reflectivity, and average 20-dBZ echo top height in the individual cells along these tracks. This approach enabled us to evaluate whether each cell underwent significant changes in structure and intensity as it passed over different regions. If a cell exhibited changes in morphology only over the urban area and remained unchanged over the other regions, it indicates

that the urban region could have an impact on the cell structure and properties. A t-test statistic was used to assess the statistical significance of the observed changes at the 95% significance level.

4.1.4.1 Upwind to Urban Region

A total of 1,367 common cell tracks were identified between the upwind and urban regions (Table 4.3), representing 13% and 18% of the total cell tracks in these regions, respectively. These tracks consisted of 6,681 cells in the upwind region and 3,438 cells in the urban region. As shown in Table 4.4, the changes in average reflectivity and average 20-dBZ echo top height were negligible as cells moved between the regions. The mean differences in these variables were 0.04 dBZ and -0.01 km, respectively, with corresponding p-values of 0.50 and 0.55, indicating that these changes were not statistically significant. Although the average cell area slightly increased by 0.49 km² as cells moved from the upwind to the urban region, the t-test-based significance test yielded a p-value of 0.22, which is above the 0.05 threshold for statistical significance. This suggests that the observed increase in cell area was not consistent enough to be considered a significant change.

Table 4.3: Distribution of common tracks and corresponding cells in the upwind and urban areas

Region	No. of tracks	No. of cells	Avg. cell area (km ²)	Avg. reflectivity (dBZ)	Avg. 20-dBZ echo top height (km)
Upwind	1,367	6,681	45.63	38.10	7.11
Urban	1,367	3,438	41.99	37.45	6.99

The lack of statistically significant changes in cell properties as they move from the upwind to the urban region implies that the initial transition into the urban environment may not have a strong immediate impact on cell structure. This finding is somewhat surprising, given the expected influence of the UHI effect and surface roughness changes on convective processes. One possible explanation is that the cells entering the urban region may take some time to respond to the surface conditions in the region, and changes in their properties may become more evident as they continue to move through the region. Another possibility is that the sample size of common cell tracks between the upwind and urban regions (1,367) may not be large enough to capture the full range of variability in cell responses to the urban environment.

Table 4.4: Summary statistics of changes in cell properties along the tracks in upwind and urban regions. Negative values indicate that cells decrease as they move from the upwind to urban region and vice versa.

	Differences in cell area (km ²)	Differences in reflectivity (dBZ)	Differences in 20-dBZ echo top height (km)
Mean	0.49	0.04	-0.01
Standard Deviation	58.93	8.39	3.27
Minimum	-789.75	-36.65	-17.00
25% Percentile	-10.75	-4.45	-1.50
50% Percentile	0	0.05	0
75% Percentile	11.25	4.54	1.50
Maximum	800.75	33.37	17.50
Statistic	-1.2	-0.67	0.60
P-value	0.22	0.50	0.55

4.1.4.2 Urban to Downwind Region

In contrast to the upwind–urban transition, the analysis of common cell tracks between the urban and downwind regions revealed statistically significant changes in cell properties. A total of 952 common cell tracks were identified, with 3,817 individual cells in the urban region and 2,699 in the downwind region. This reduction in the number of cells could indicate cell merging or dissipation as they move into the downwind area.

Table 4.5: Distribution of common tracks and corresponding cells in the urban and downwind areas

Region	No. of tracks	No. of cells	Average cell area (km ²)	Average reflectivity (dBZ)	Average 20-dBZ echo top height (km)
Urban	952	3,817	51.44	39.55	7.62
Downwind	952	2,699	45.39	38.38	7.31

The average cell area decreased from 51.44 km² in the urban region to 45.39 km² in the downwind region, with the cells in each cell track decreasing by an average of 2.95 km². The t-statistic (4.87) and p-value (1×10^{-6}) indicate that this decrease in cell area is statistically significant. Similarly, both the average reflectivity and echo top height decreased by 0.50 dBZ (p-value = 5.42×10^{-10}) and 0.08 km (p-value = 1.5×10^{-2}), respectively, as the cells moved into the downwind areas. Although these changes were small, they were statistically significant at the 95% confidence interval.

The significant decreases in cell area, reflectivity, and echo top height as cells move from the urban to the downwind region suggest that the urban environment has a considerable impact on the characteristics of cells that move towards the downwind area. These findings are consistent with the idea that the urban region can influence the evolution and dissipation of convective cells (Niyogi et al., 2011; Shepherd et al., 2010). The decrease in cell area could be attributed to the dissipation of cells as they

encounter different surface and atmospheric conditions in the downwind region. The reductions in reflectivity and echo top height indicate a weakening intensity and a decrease in the vertical extent of the cells, respectively. These changes may be related to the disruption of the cells' movement as they move away from the UHI and encounter different surface roughness and environmental conditions in the downwind area.

Table 4.6: Summary statistics of changes in cell properties in urban and downwind regions. Negative values indicate that cells decrease as they move from urban to downwind region and vice versa.

	Differences in cell area (km ²)	Differences in maximum reflectivity (dBZ)	Differences in maximum 20-dBZ echo top height (km)
Mean	-2.95	-0.50	-0.08
Standard Deviation	62.38	8.25	3.31
Minimum	-744.75	-39.06	-12.50
25% Percentile	-14.00	-4.85	-2.00
50% Percentile	-0.50	-0.43	0
75% Percentile	10.75	3.85	2.00
Maximum	578.75	34.08	13.50
Statistic	4.87	6.21	2.43
P-value	1×10^{-6}	5.42×10^{-10}	1.5×10^{-2}

4.2 Diurnal Variations in Convective Cell Development

Over the entire study region, the number of new cells increases significantly through the morning hours and peaks at 2 PM (Fig. 4.7), which is the time of maximum surface

heating. As surface temperatures rise throughout the day, atmospheric conditions become increasingly conducive to convection initiation. The dissipation of these cells follows a similar diurnal pattern, but they lag behind by a few hours and peak at 4 PM local time. This delay suggests that once triggered, the cells typically persist for about 2-4 hours before dissipating. The sharp decline in end tracks after 10 PM indicates that most cells dissipate by late evening, which coincides with the cooling of the surface and the stabilization of the lower atmosphere.

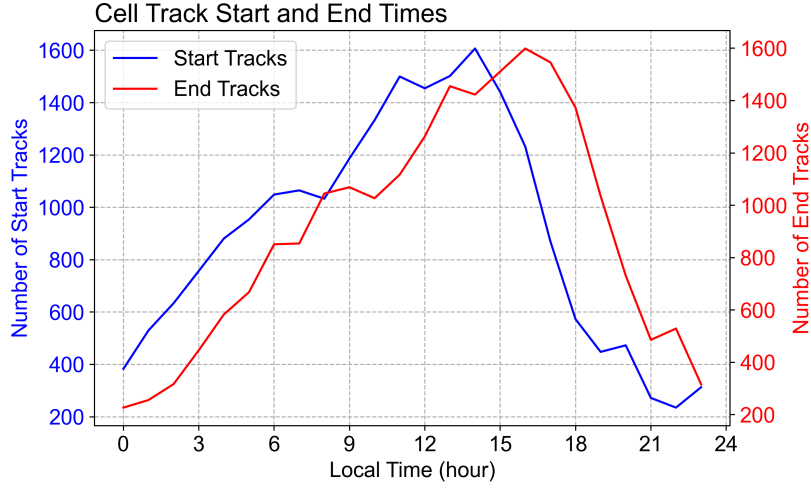


Figure 4.7: Diurnal distribution of start and end tracks of the cells

4.2.1 Diurnal Variations in the Upwind Region

In the upwind region, the diurnal cycle of cell area exhibits some variability, with an initial peak of approximately 33 km^2 at about 2 AM, followed by a decrease to approximately 23 km^2 by 5 AM (Fig. 4.8). Subsequently, the cell area rapidly increases again, reaching nearly 33 km^2 by midday (12-1 PM local time). The slight increase in maximum reflectivity and echo top height coincides with the increasing cell area in the morning hours, suggesting that the cells are growing in both size and intensity during

this period. However, both variables remain relatively constant throughout the day, with a slight decrease and higher variability observed during the evening hours.

Interestingly, the result reveals that almost 80% of the total upwind cells are initiated in the early morning hours, steadily reducing to a low of 25% through the afternoons before increasing again into the night (Figs. 4.2 and 4.9). This higher percentage of early morning initiations is attributed to the presence of nocturnal land breezes and low-level convergence, which act as triggers for convective activity. The subsequent reduction in the afternoon may be due to the suppression of convection over the cooler sea surface and the inland propagation of the sea breeze front, which pushes the cells further inland (Johnson, 2011). The nighttime increase in the percentage of upwind cells initiated, despite the decreasing cell area, could be due to the re-establishment of land breezes that promote initiation, though with smaller cell sizes compared to those that develop during the afternoon hours, consistent with the findings of Niyogi et al. (2011). While cell initiation is primarily driven by the sea breeze in the early morning, moderate increases are observed as diurnal heating progresses. The lower overall initiation rates during the day in this region, compared to the urban and downwind regions, reflect the influence of the sea breeze rather than localized heating.

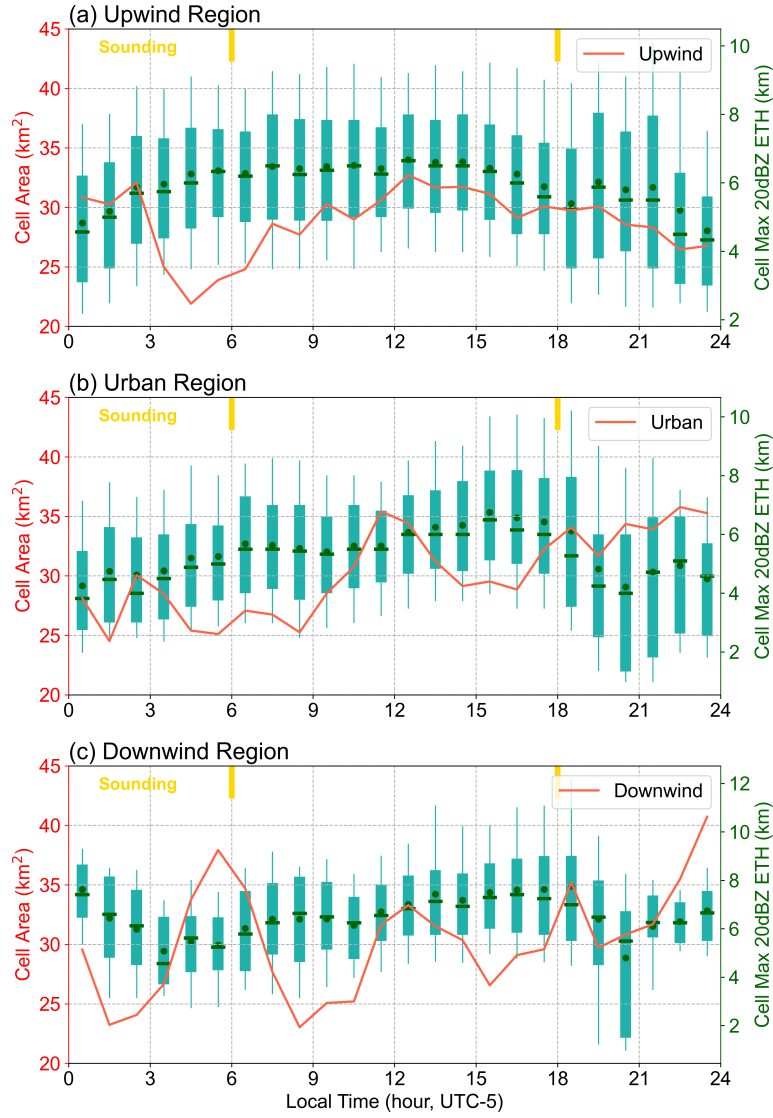


Figure 4.8: Diurnal variation of cell area and maximum 20-dBZ echo-top height in the upwind, urban, and downwind regions. The thick line represents the median, while the dot indicates the mean. The whiskers show the range, and they extend from the 10th percentile to the 90th percentile to exclude potential outliers beyond these limits.

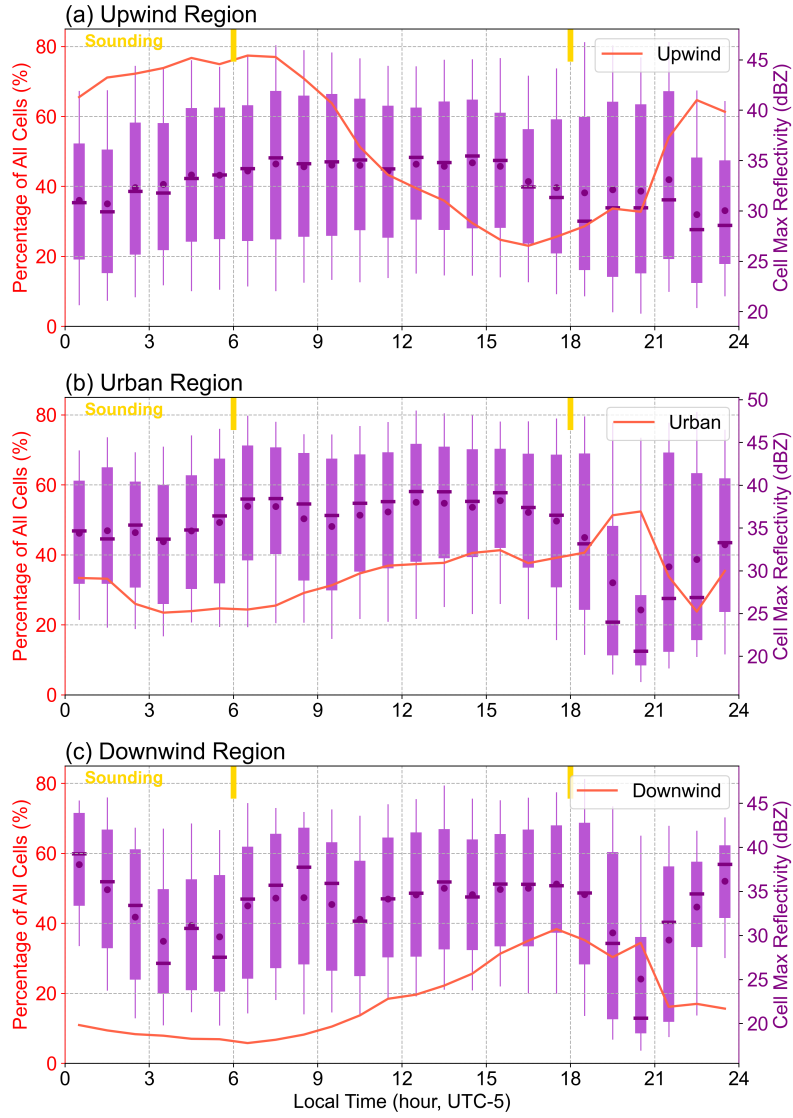


Figure 4.9: Diurnal variation of the percentage of all cells initiated and the maximum reflectivity in the upwind, urban, and downwind regions. The thick line represents the median, while the dot indicates the mean. The whiskers show the range, and they extend from the 10th percentile to the 90th percentile to exclude potential outliers beyond these limits.

4.2.2 Diurnal Variations in the Urban Region

In the urban region, the cell area exhibits more variability in the early morning hours compared to the upwind region, gradually increasing to maximum of around 30 km² at 2 AM and 35 km² at 11 AM (Fig. 4.8). Although the cell area fluctuates in the afternoon, it shows an upward trend again, reaching 35 km² by 10 PM. The percentage of cells initiated in this region remains relatively stable throughout the day, fluctuating between 30–40%. It increased during the afternoon and evening hours (12-21 hours local time), and reached a maximum of around 40% (Figs. 4.2 and 4.9). The maximum echo top height also gradually increases from the early morning but begins to decrease after 7 PM, exhibiting higher variability during this period. A similar trend is observed for reflectivity, with a pronounced dip and larger variability during the evening hours, although the overall values remain relatively constant at approximately 38 dBZ for most part of the day.

This region also experiences the highest initiation activity during the peak heating period from 10 AM to 4 PM (Fig. 4.2), driven by the UHI effect. Tuftedal et al. (2023) found similar results in Houston and attributed it to the advancing sea breeze front from the upwind region. The sustained initiation into the early evening hours highlights the significant impact of urban heating on convection processes. The decreasing echo top heights and higher reflectivity variability in the evening also suggest a transition to more localized and variable convective activity as the UHI effect weakens. Notably, the UHI effect sustains convective activity into the late afternoon, while cooling in the evening leads to the termination of cells.

4.2.3 Diurnal Variations in the Downwind Region

The downwind region exhibits the most complex diurnal cycle in cell area among the regions, with multiple peaks observed throughout the day and night. The cell area begins to increase in the early morning, reaching a peak of about 38 km² around 5 AM before decreasing (Fig. 4.8). It then increases again in the mid-morning, reaching another peak by 1 PM. Additional peaks in cell area are observed at 7 PM and at midnight, suggesting a multimodal distribution in the downwind region. In contrast to the upwind and urban regions, less than 20% of the total cells are initiated in the downwind region during the morning hours. However, this percentage begins to increase in the late afternoon to night (18-24 hours) and reaches a maximum of about 40%, indicating that the majority of cells in the downwind region are initiated during the late evening and nighttime hours (Figs. 4.2 and 4.9).

The maximum reflectivity and echo top height trends in the downwind region follow patterns similar to those observed in the urban region, suggesting that the convective cells exhibit comparable vertical development and intensity. The multimodal peaks in cell area and the delayed increase in the percentage of cells initiated in this region may be attributed to several factors: (1) the inland propagation of convective systems from the upwind and urban regions by the sea breeze (Tuftedal et al., 2023), (2) the interaction of the sea breeze front with other mesoscale boundaries, such as outflow boundaries from previous convection, triggering new cell initiation or intensification, or (3) the advection of moisture and instability from the upwind regions, creating a more favorable environment for convection initiation and growth as the cells propagate inland.

4.2.4 Effect of Distance From the Urban Edge

In the upwind region, the average cell area increases from approximately 43 km² to around 52 km² as the distance to the urban edge decreases from 50 km (Fig. 4.10). This increase suggests that the cells initiated near the urban boundary tend to be larger in size and are potentially influenced by sea breezes in the upwind region, which contributes to the variability in their dimensions. Furthermore, the average reflectivity also increases from about 37 dBZ to 43 dBZ, indicating that these cells become more intense towards the urban area. However, this intensification is accompanied by a decrease in the average echo top height, decreasing from approximately 7.5 km to 6.5 km towards the urban edge. This decrease suggests that while the cells gain intensity near the urban region, their vertical development is constrained, potentially due to the complex surface features and localized convection patterns in the urban environment.

In the downwind region, the trends in cell properties exhibit a more nuanced behavior. The average cell area initially decreases with increasing distance from the urban edge, starting at around 45 km² before dipping to approximately 41 km². However, as the distance from the urban edge increases further, the cell area begins to increase again, reaching approximately 45 km². Notably, the cell area in the downwind region is generally lower than that in the upwind region at the urban edge (45 km² vs. 52 km²), indicating that the cells may experience an initial increase in size over the urban area due to the UHI effect. The average reflectivity in the downwind region follows an increasing trend, starting at about 37 dBZ near the urban edge and reaching approximately 40 dBZ as the distance increases, with variability observed in the shaded area. This increase suggests that cells intensify as they propagate inland, with some cells remaining weaker and others becoming quite intense. Moreover, the average echo top height in the downwind increases from approximately 8.5 km near the urban edge to around 10 km as the distance increases. This indicates that the cells grow vertically

as they move inland, with the higher echo tops further from the urban edge suggesting sustained and potentially merging process taking place. Collectively, the trends in this region point towards sustained and merging convective activity, leading to the development of larger and more intense cells as they propagate further inland and away from the influence of the urban area. While the downwind region exhibits evidence of intensifying convective cells, this effect is generally smaller in magnitude compared to the urban region (Fig. 4.5).

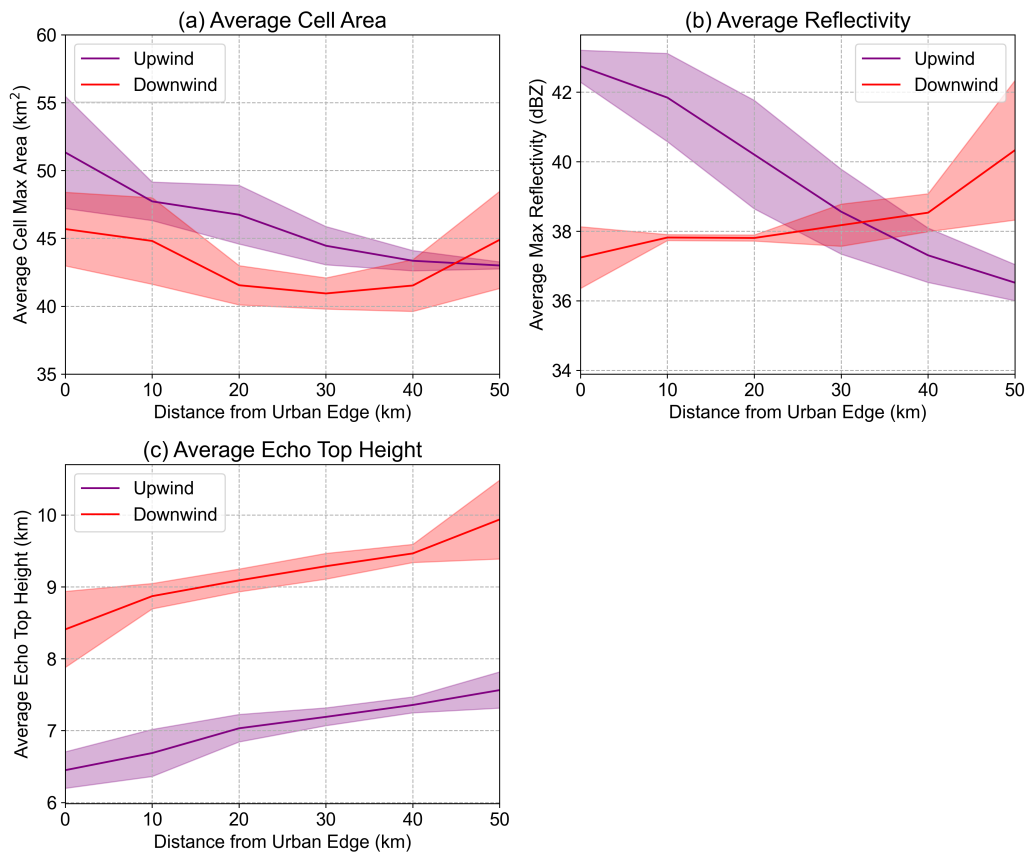


Figure 4.10: Running mean and standard deviation of the average (a) cell area, (b) reflectivity, and (c) echo top height at 20 dBZ in each region by distance from the edge of the urban area

4.3 Characterizing Cell Properties across Lifetimes

To gain a comprehensive understanding of the spatiotemporal evolution of convective cells in the Houston area, it is crucial to examine their properties across different lifecycle stages. This section focuses on the developing, mature, and dissipating stages of the cells within the upwind, urban, and downwind regions.

4.3.1 Developing Stage

The developing stage marks the initial growth phase of convective cells, where they begin to form and intensify. According to Fig. 4.11, this stage has the highest proportion of cells compared to the mature and dissipating stages across all regions throughout the day (Fig. 4.11). However, the diurnal patterns of cell initiation vary considerably between the regions, with the upwind region experiencing a higher proportion of developing cells in the early morning and afternoon hours (Fig. 4.11(a,c)), while the urban and downwind regions show a more gradual increase from morning to afternoon (Fig. 4.11(b,c)). The early morning peak in developing cells in the upwind region aligns with the increased cell initiations driven by the intensifying sea breeze circulation during the night-to-day transition (Figs. 4.2 and 4.9). This supports our previous findings that the land-sea temperature contrast plays a crucial role in triggering convection along the coast. Meanwhile, the upwind region maintains a slightly higher proportion of developing cells in the afternoon than the other regions (Fig. 4.11(c)), despite the expected increase in the urban and downwind regions due to the influence of daytime heating and UHI effects. This could be attributed to the continuous triggering of new cells along the sea breeze front as it propagates inland, a mechanism that may be more persistent and dominant in the upwind region compared to the urban and downwind areas (Fovell, 2005).

The cell properties during this stage further reveal the characteristics of each region. The upwind region exhibits moderate average cell areas, reflectivity values, and echo top heights (Fig. 4.12), indicating that the initial convection growth is primarily influenced by sea breeze convergence and limited instability. In contrast, the urban region shows larger average cell areas and higher average reflectivity values (Fig. 4.12(a,b)), suggesting that the UHI effect and advancing sea breeze front both enhance the formation and early growth of cells in this region. However, the similarity in mean echo top heights between the urban and upwind regions during the developing stage (Fig. 4.12(c)) raises an interesting question: why does the UHI effect not significantly enhance the average vertical development of cells in the urban region in the early stages? This could be due to several factors, such as the limited time for cell to mature within the urban environment, the presence of turbulent mixing, or increased surface roughness that may limit vertical growth. It is also possible that the UHI effect primarily enhances the horizontal growth of developing cells, as indicated by the larger cell area in the urban region compared to the upwind area (Fig. 4.12(a)), while the vertical development may be more strongly influenced by other factors, such as the advection of instability from the upwind regions. The enhanced vertical development in the downwind region suggests that the advection of moisture and instability from the upwind areas, combined with the potential extension of the UHI effect, creates a favorable environment for cell growth and allows cells to reach greater altitudes during their early stages.

Meanwhile, the average cell duration in the developing stage varies across the regions, with cells in the urban region spending an average of 23 minutes in this stage before developing into the mature stage, followed by the upwind region with 21 minutes, and the downwind region with 20 minutes (Fig. 4.13). The longer duration in the urban region may allow sufficient time for the cells to grow and organize due to the

influence of the UHI effect. However, this does not necessarily imply that the urban cells have a longer overall lifetime, as observed in Fig. 4.5. If the surrounding environment of the cells lacks sufficient moisture supply to sustain them, their overall lifetime can be limited even if they spend a prolonged time in the developing stage (Fovell and Tan, 1998). In contrast, the upwind region may have a higher moisture supply from the Gulf, which can help sustain the cells for a longer period, even after a relatively short developing stage (Tuftedal et al., 2023).

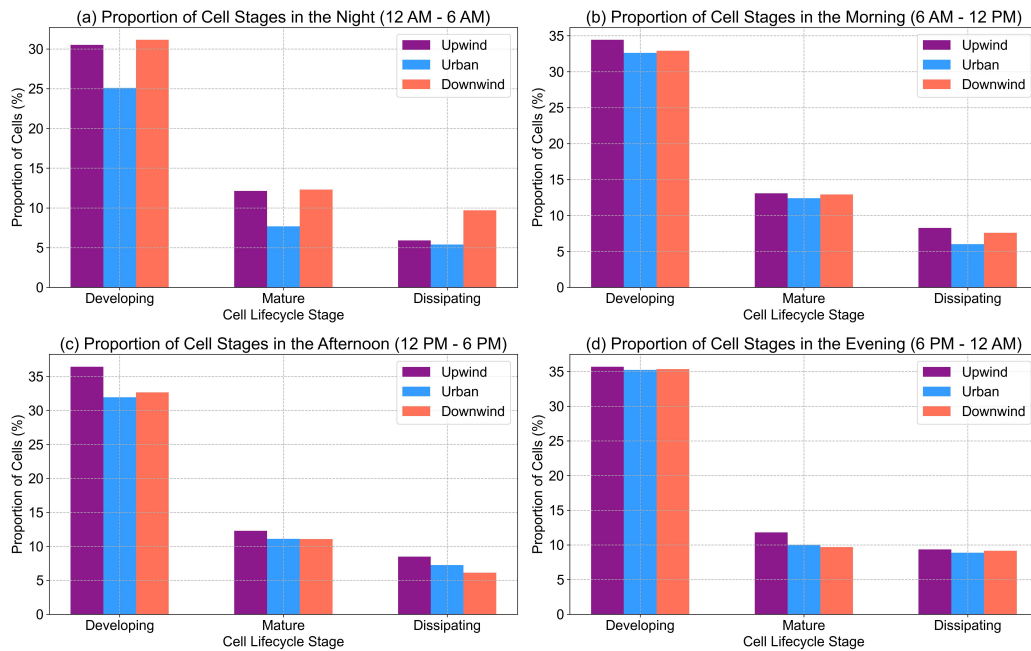


Figure 4.11: Diurnal variation in proportion of cell lifecycle stages across each region

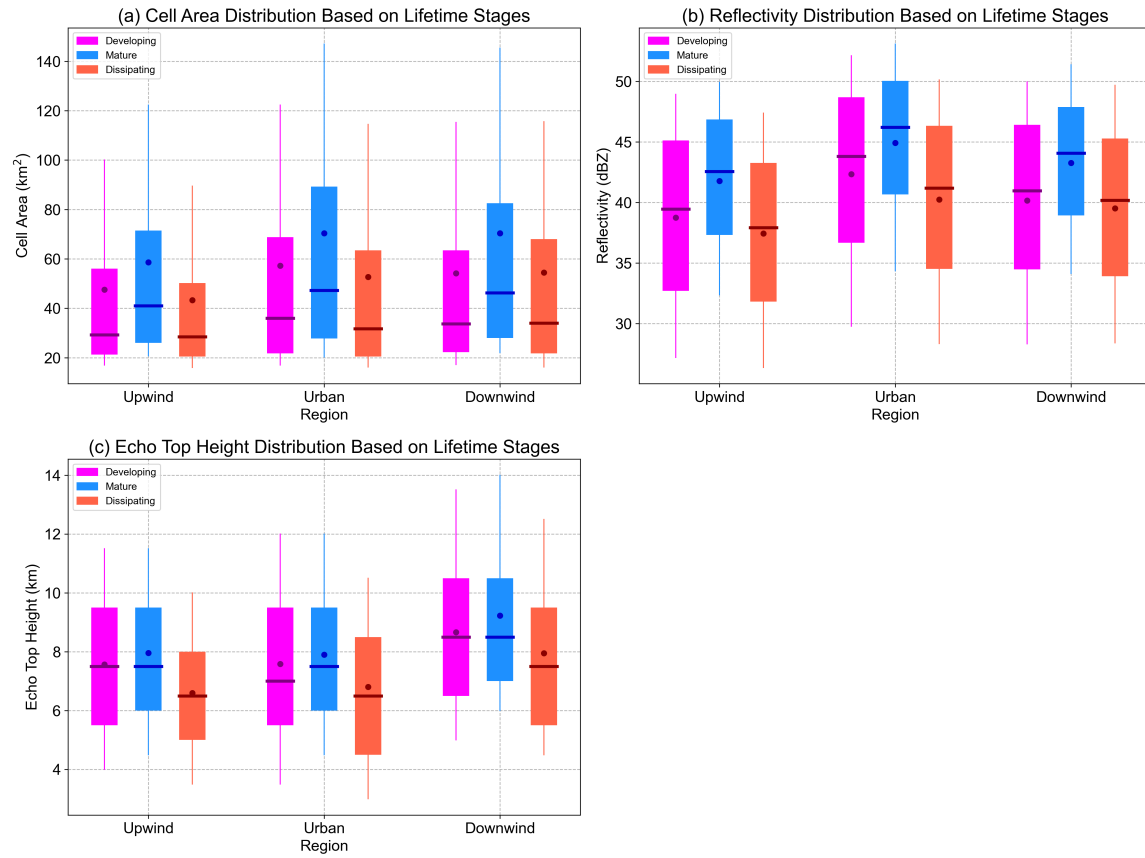


Figure 4.12: Distribution of cell properties across different regions based on lifetime stages. (a) Average cell area distribution, (b) Maximum reflectivity distribution, and (c) Maximum 20-dBZ echo top height distribution. The thick line represents the median, the dot indicates the mean, and whiskers show the range.

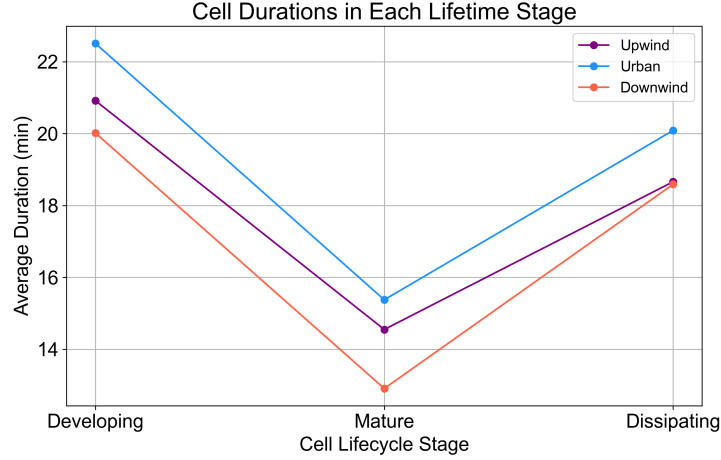


Figure 4.13: Average duration of cell lifecycle stages in different regions

4.3.2 Mature Stage

The mature stage represents the peak intensity and organization of the cells, and it is characterized by well-developed updrafts, high reflectivity values, and significant vertical growth. During this stage, cells exhibit their maximum areal extent, reflectivity, and echo top heights, indicating the presence of strong convective processes and intense precipitation. Compared to the developing stage, the results show that the proportion of cells in the mature stage is lower across all regions throughout the day (Fig. 4.11). This suggests that fewer cells reach this stage, and those that do tend to spend less time in this stage compared to the developing stage (Fig. 4.13). The diurnal occurrence of mature cells also vary between the regions, with the upwind and downwind regions showing higher proportions during the night and morning hours, while the urban region exhibits a more gradual increase in mature cell proportions from morning to afternoon. As the day progresses, the proportion of mature cells in the urban region begins to catch up with those in the upwind and downwind regions. This increase may

be influenced by the combination of daytime heating, increased surface roughness, and the interaction between the advancing sea breeze front and the urban boundary layer.

The cell properties during the mature stage reveal the peak intensity reached by cells in each region. The upwind region in this stage exhibits a significant increase in average cell area, reflectivity, and echo top height compared to those in the developing stage (Fig. 4.12), indicating the growth and intensification of cells as they mature. Notably, the downwind region exhibits the highest average echo top heights among the regions during the mature stage, reaching altitudes of 9.5 km (Fig. 4.12(c)). Furthermore, the average duration of cells in the mature stage is shorter compared to the developing stage across all regions (Fig. 4.13). The urban region has the longest mature stage duration of 15 minutes, followed by the upwind region with 14.5 minutes and the downwind region with 13 minutes. The shorter duration of the mature stage suggests that cells tend to quickly reach their peak intensity and then transition to the dissipating stage or merge with other cells. The slightly longer mature stage duration in the urban region may be related to the sustained low-level convergence by the interaction between the urban boundary layer and the advancing sea breeze front, which can support cell longevity.

4.3.3 Dissipating Stage

The dissipating stage marks the weakening and decay of the cells, characterized by a decrease in cell area, reflectivity, and echo top heights. During this stage, cells begin to lose their organization and intensity, leading to a gradual dissipation of the convective activity. The proportion of cells in the dissipating stage varies across the regions throughout the day (Fig. 4.11). The downwind region exhibits the highest proportion of dissipating cells during the night hours, followed by the upwind and urban regions. This suggests that the cells advected into the downwind region from the upwind and urban areas tend to dissipate more frequently during the nighttime, possibly due to the

absence of daytime heating and the weakening of the sea breeze circulation. As the day progresses, the proportion of dissipating cells in the urban region decreases, reaching the lowest values during the morning hours (Fig. 4.11(b)). This may be attributed to the influence of the developing UHI effect and the increased surface heating, which can sustain convective activity and delay the onset of cell dissipation. In contrast, the upwind region shows the highest proportion of dissipating cells during the afternoon hours, followed by the urban and downwind regions (Fig. 4.11(c)).

The cell properties during the dissipating stage reveal the largest decrease in cell area, lowest reflectivity values, and lowest echo top heights in the upwind region compared to the urban and downwind regions (Fig. 4.12(a)). These properties indicate that the upwind cells generally tend to dissipate more rapidly and have a less organized structure during the dissipating stage. While sea breeze forcing is a dominant feature in this region, its effect appears to be overshadowed by the absence of strong forcing mechanisms compared to the urban and downwind areas. The lack of stronger forcing mechanisms, such as UHI or convergence zones, likely contributes to the faster dissipation of cells in this region. Meanwhile, the values of the cell properties were slightly higher in the urban region during the dissipating stage (Fig. 4.12). The presence of the UHI effect and the interaction with the sea breeze front may help to sustain the cells in the urban region for a slightly longer duration compared to the upwind area. However, the gradual decrease in surface heating and the weakening of the UHI effect during the late afternoon and evening hours can lead to the dissipation of cells in this region. In contrast to the rapid dissipation observed in the upwind region, the downwind area exhibits a remarkable persistence of convective cells during the dissipating stage. The data reveals that the downwind region exhibits the largest average cell area, highest reflectivity values, and highest echo top heights during this stage (Fig. 4.12). These characteristics suggest that cells in the downwind region maintain a more organized

structure and intensity even as they begin to weaken. The accumulation of moisture and instability in the downwind region, as well as the potential merging of cells advected from the upwind and urban areas, may contribute to the slower dissipation of cells in this region.

Overall, the analysis of the cell properties across different lifecycle stages reveals distinct patterns in different regions. The developing stage shows higher proportions of cells in the upwind region during early morning and afternoon hours, likely driven by sea breeze circulation. The urban region exhibits larger cell areas and higher reflectivity values in this stage, suggesting UHI effects enhance early cell growth. The mature stage sees shorter durations but peak intensities across all regions, with the downwind region showing the highest echo top heights. The dissipating stage highlights the persistence of cells in the downwind region compared to rapid dissipation upwind.

To further understand the factors driving these observed differences in cell characteristics and evolution, it is crucial to examine the near-storm environments across the three regions. The following section compares key environmental parameters between the upwind, urban, and downwind areas to elucidate how variations in atmospheric conditions influence convective cell behavior throughout their lifecycles.

4.4 Comparing Near-storm Environments across Regions

4.4.1 Upwind vs Downwind Regions

Using ARM INTERPSONDE data, we compared various environmental metrics between the upwind and downwind regions (Table 4.7). The analysis revealed substantial

differences in the near-storm environments between both regions, and these differences were statistically significant at the 95% confidence level for several key parameters.

In the downwind region, the CAPE (2170 J/kg) was significantly higher than that in the upwind region (2081 J/kg). This enhanced instability parameter indicates a more favorable environment for intense thunderstorm development in the downwind region compared to the upwind region. Higher CAPE values can lead to stronger updrafts, which can lift moisture higher into the atmosphere, resulting in deep clouds. The increased instability allows for more efficient precipitation processes, leading to higher rainfall totals per cell. The influence of daytime solar heating in the downwind region likely contributes to the enhanced instability, in addition to boundary layer mixing and entrainment of warmer air aloft. This amplified instability aligns with the characteristics of cells in this region, as observed in previous sections. The downwind area exhibited some of the highest echo-top heights among the regions and a higher total average precipitation per cell (30 mm) compared to the upwind region (20 mm). This suggests that the unstable environment may allow the cells to intensify and produce more efficient precipitation as they propagate inland. In contrast, the lower CAPE in the upwind region limits the intensity and vertical growth of cells initiated in this region, as reflected by the relatively smaller average maximum cell area (42.5 km²), moderate average maximum 20-dBZ echo top height (8 km), and lower total precipitation per cell (Fig. 4.5). The higher most unstable (MU) CAPE value in the downwind region also aligns with the observed larger echo top height, suggesting that the cells may be able to achieve significant vertical development through a deeper tropospheric layer owing to greater potential instability. While low-level moisture (850-hPa dewpoint temperature and relative humidity) was slightly reduced in the downwind region compared to the upwind region, the high CAPE appeared to

compensate, allowing robust precipitation production as evident from the highest total rainfall per cell (30 mm) occurring downwind among the three regions analyzed.

The downwind region also exhibited slightly modified low-level wind shear and helicity profiles compared to upwind, though the differences were minimal. The 0–1 km storm-relative helicity was $0.12 \text{ m}^2/\text{s}^2$ downwind, contrasting with $-3.11 \text{ m}^2/\text{s}^2$ upwind. While these values indicate a shift from slightly negative to slightly positive helicity, both are extremely low and unlikely to significantly impact mesocyclone development on their own. The 0–1 km bulk shear was marginally lower downwind at 4.17 m/s versus 4.42 m/s upwind, a difference that is likely negligible for storm organization. At deeper layers, the 0–6 km bulk shear was modestly lower downwind (5.64 m/s) compared to upwind (5.86 m/s). This decrease in deep-layer shear, though small, aligns with the observed weakening of upper-level winds downwind.

The wind profile exhibited slight but statistically significant differences between the upwind and downwind areas. Surface winds were modestly stronger downwind at 5.10 m/s compared to 4.57 m/s upwind, while 850-hPa winds showed a small increase from 5.08 m/s upwind to 5.32 m/s downwind. However, the upper-level 500 hPa winds exhibited a minor decrease, with 3.02 m/s downwind versus 3.22 m/s upwind. While these wind speed differences were relatively small, the resulting pattern suggests the development of a slightly enhanced low-level jet and a modest weakening of upper-level flow downwind. The modest increase in low-level winds downwind may have aided in sustaining and advecting convective cells as they moved inland from the urban region. This subtle low-level jet enhancement, combined with the significantly higher instability, could contribute to more organized updrafts and increased cell longevity as observed in the downwind region. Meanwhile, the minor decrease in upper-level winds downwind may have reduced ventilation and storm-relative flow, potentially allowing

cells to persist without being as readily disrupted or advected away from the area compared to the upwind region.

It is important to note that while the shear and helicity values alone do not suggest a high likelihood of organized, rotating storms, the combination of slightly enhanced low-level winds and significantly increased instability downwind may have contributed to the observed characteristics of convective cells in this region, including higher echo-top heights and increased precipitation per cell. The observed differences in cell behavior between the upwind and downwind regions are likely driven more by the substantial increase in CAPE rather than the minimal changes in shear and helicity values.

4.4.2 Upwind vs Urban Regions

The near-storm environments between the upwind and urban regions reveal important differences in atmospheric conditions that influence their convection initiation, organization, and intensity (Table 4.8). These differences are reflected in the empirical distributions and characteristics of cells in these two regions.

The urban region exhibited significantly higher CAPE values at the surface (2125.13 J/kg) and most unstable levels (2204.74 J/kg) compared to the upwind region (2081.06 J/kg and 2169.92 J/kg, respectively). These higher CAPE values indicate a greater amount of potential energy available for air parcels to accelerate upwards and fuel convective updrafts. While the temperature differences between the urban and upwind regions were relatively small at the mixed-layer (27.48°C urban vs. 27.19°C upwind) and most unstable levels (28.76°C urban vs. 28.32°C upwind), these elevated temperatures, likely resulting from the UHI effect, played a contributing role in increasing the buoyancy of air parcels in the urban area. However, the primary driver for the

enhanced instability and convective initiation in this region appears to be the substantially higher CAPE values, rather than the modest temperature differences alone. The higher CAPE values, coupled with the UHI-induced destabilization and localized hotspots in the urban area, provided a more favorable environment for air parcels to overcome the convective inhibition and release their buoyant energy, leading to the observed higher frequency of convective cell initiation events and the high cell density of 20 cells per 10 km² in the urban region compared to the upwind area (see Section 4.1).

While the moisture content in the urban region showed mixed signals, with slightly lower relative humidity at the 500 hPa (79.68% urban vs. 80.34% upwind) and 850 hPa levels (66.43% urban vs. 67.41% upwind), the most unstable dewpoint temperature was marginally higher (23.30°C urban vs. 23.33°C upwind). These observations suggest that the low-level moisture content may have been slightly lower in the urban area compared to the upwind region. However, the enhanced instability and UHI-induced destabilization likely played a more dominant role in supporting convection initiation and development, potentially compensating for the slightly drier low-level conditions.

The wind parameters and shear profiles also exhibited differences between the upwind and urban regions, though their impact on convective cell organization and severity may be limited. The storm-relative helicity values at the 0–1 km ($-1.35 \text{ m}^2/\text{s}^2$), 0–3 km ($10.25 \text{ m}^2/\text{s}^2$), and 0–6 km ($21.18 \text{ m}^2/\text{s}^2$) levels were higher in the urban region, as were the bulk shear values at the 0–1 km (4.32 m/s) and 0–3 km (5.08 m/s) levels. While these values are higher than in the upwind region, they are still relatively low and may not significantly impact low-level rotation or storm organization on their own. The observed larger cell areas (45.5 km²) and higher reflectivity values (39.5 dBZ) in the urban region may be more strongly influenced by the enhanced instability rather

than these modest increases in shear and helicity. Furthermore, the urban region experienced slightly stronger surface winds (4.70 m/s urban vs. 4.57 m/s upwind) and lower wind speeds at the 500 hPa (3.04 m/s urban vs. 3.22 m/s upwind) and 850 hPa levels (5.29 m/s urban vs. 5.08 m/s upwind). While this wind profile, with stronger low-level winds and weaker upper-level winds, could potentially contribute to low-level convergence and lift, the differences are small and their impact on convection initiation and sustenance may be limited.

The combination of these factors – primarily the significantly higher CAPE values with possible minor contributions from the wind profile – may have contributed to the higher variability in precipitation patterns observed in the urban region, with occasional intense rainfall events occurring alongside periods or areas of less intense precipitation. However, it should be noted that the observed differences in convection initiation, organization, and precipitation patterns are likely driven predominantly by the substantial increase in instability parameters rather than the modest variations in wind profiles and shear values. Further research would be needed to quantify the relative contributions of these factors to the observed convective behavior in the urban environment.

4.4.3 Urban vs Downwind Regions

The downwind region exhibited slightly higher CAPE values at the surface (2169.75 J/kg vs. 2125.13 J/kg in the urban region) and most unstable levels (2239.38 J/kg vs. 2204.74 J/kg) (Table 4.9), indicating a marginally more unstable environment compared to the urban region. However, the differences in CAPE values were relatively small, suggesting that both regions were characterized by high instability. Interestingly, despite slightly lower CAPE values, the urban region exhibited higher cell densities (20

cells per 10 km²), larger cell areas (45.5 km²), and higher reflectivity values (39.5 dBZ) compared to the downwind region. This suggests that other factors, such as the UHI effect, may have played a significant role in influencing the convective cell characteristics in the urban region.

The moisture content showed minor variations, with slightly lower dewpoint temperatures in the downwind region at the surface (68.52°C vs. 68.74°C in the urban region) and 500 hPa level (79.59°C vs. 79.68°C), while the most unstable dewpoint temperature showed no significant difference between the two regions. The wind shear and storm-relative helicity profiles revealed some differences between the two regions, though their magnitudes were generally low. The downwind region exhibited slightly higher storm-relative helicity at the 0–1 km (0.12 m²/s² vs. -1.35 m²/s² in the urban region), 0–3 km (11.57 m²/s² vs. 10.25 m²/s²), and 0–6 km levels (19.72 m²/s² vs. 21.18 m²/s²). While these values are higher than in the urban region, they are still relatively low and may not significantly impact low-level rotation or storm organization on their own. The downwind region had slightly higher bulk shear values at the 0–1 km (4.17 m/s vs. 4.32 m/s in the urban region), 0–3 km (5.42 m/s vs. 5.08 m/s), and 0–6 km levels (5.64 m/s vs. 5.89 m/s). These modest increases in wind shear throughout the lower and mid-troposphere may have contributed to some extent to the organization and longevity of the cells as they propagated into the downwind region, but their impact should not be overstated given the relatively low magnitudes. The wind profile showed that the downwind region experienced stronger surface winds (5.10 m/s vs. 4.70 m/s in the urban region) but comparable wind speeds at the 500 hPa and 850 hPa levels. This slightly stronger low-level flow in the downwind region may have played a minor role in sustaining and advecting convective cells as they moved inland from the urban area.

While the empirical observations noted that some cells persisted up to 5 hours in the downwind region, this longevity is likely due to a combination of factors, including the high instability, rather than solely the modest increases in wind shear and helicity. The downwind environment appears to be conducive to the organization, vertical development, and precipitation efficiency of these cells, driven primarily by the high instability, with potential minor contributions from the slightly enhanced wind shear and marginally higher helicity values.

Table 4.7: Summary statistics of changes in cell properties in upwind and downwind regions. ML = mixed layer, MU = most unstable layer. Both MU and ML are within the lowest 50 hPa.

Environmental metric	Upwind (mean)	Downwind (mean)	p-value	Significant?
CAPE (J/kg)	2081	2170	1.36×10^{-13}	Yes
CIN (J/kg)	-36	-36	0.4	No
ML Temperature ($^{\circ}\text{C}$)	27.2	27.7	1.22×10^{-79}	Yes
ML Dewpoint Temperature ($^{\circ}\text{C}$)	22.5	22.4	1.30×10^{-7}	Yes
ML CIN (J/kg)	-4	-4	0.1	No
MU Temperature ($^{\circ}\text{C}$)	28.3	29.2	2.07×10^{-110}	Yes
MU Dewpoint Temperature ($^{\circ}\text{C}$)	23.3	23.3	0.7	No
MU CAPE (J/kg)	2170	2239	7.51×10^{-10}	Yes
MU CIN (J/kg)	-28	-32	4.64×10^{-9}	Yes
0-1km Storm Relative Helicity (m^2/s^2)	-3.1	0.1	8.14×10^{-51}	Yes
0-3km Storm Relative Helicity (m^2/s^2)	8.6	11.6	4.08×10^{-31}	Yes
0-6km Storm Relative Helicity (m^2/s^2)	20.3	19.7	2.36×10^{-2}	Yes
0-1km Bulk Shear (m/s)	4.4	4.2	8.43×10^{-13}	Yes
0-3km Bulk Shear (m/s)	4.9	5.4	2.95×10^{-29}	Yes
0-6km Bulk Shear (m/s)	5.9	5.6	3.55×10^{-6}	Yes
Surface Wind Speed (m/s)	4.6	5.1	1.39×10^{-69}	Yes
500-hPa Wind Speed (m/s)	3.2	3.0	0	Yes
850-hPa Wind Speed (m/s)	5.1	5.3	0	Yes
500-hPa Relative Humidity (%)	80.3	79.6	0	Yes
850-hPa Relative Humidity (%)	67.4	66.3	0	Yes
Surface Dewpoint Temperature ($^{\circ}\text{C}$)	70.4	68.5	0	Yes
500-hPa Dewpoint Temperature ($^{\circ}\text{C}$)	80.3	79.6	0	Yes
850-hPa Dewpoint Temperature ($^{\circ}\text{C}$)	67.4	66.3	0	Yes

Table 4.8: Summary statistics of changes in cell properties in upwind and urban regions. ML = mixed layer, MU = most unstable layer. Both MU and ML are within the lowest 50 hPa.

Environmental metric	Upwind (mean)	Urban (mean)	p-value	Significant?
CAPE (J/kg)	2081	2125	1.28×10^{-5}	Yes
CIN (J/kg)	-36	-34	3.49×10^{-3}	Yes
ML Temperature ($^{\circ}\text{C}$)	27.2	27.5	2.32×10^{-38}	Yes
ML Dewpoint Temperature ($^{\circ}\text{C}$)	22.5	22.4	1.85×10^{-3}	Yes
ML CIN (J/kg)	-4	-3	1.95×10^{-9}	Yes
MU Temperature ($^{\circ}\text{C}$)	28.3	28.8	2.24×10^{-40}	Yes
MU Dewpoint Temperature ($^{\circ}\text{C}$)	23.3	23.3	3.15×10^{-2}	Yes
MU CAPE (J/kg)	2170	2205	3.39×10^{-4}	Yes
MU CIN (J/kg)	-28	-28	0.983	No
0-1km Storm Relative Helicity (m^2/s^2)	-3.1	-1.4	3.35×10^{-21}	Yes
0-3km Storm Relative Helicity (m^2/s^2)	8.6	10.3	4.39×10^{-15}	Yes
0-6km Storm Relative Helicity (m^2/s^2)	20.3	21.2	6.95×10^{-6}	Yes
0-1km Bulk Shear (m/s)	4.4	4.3	1.06×10^{-3}	Yes
0-3km Bulk Shear (m/s)	4.9	5.1	1.05×10^{-6}	Yes
0-6km Bulk Shear (m/s)	5.9	5.9	0.361	No
Surface Wind Speed (m/s)	4.6	4.7	3.36×10^{-7}	Yes
500-hPa Wind Speed (m/s)	3.2	3.0	0	Yes
850-hPa Wind Speed (m/s)	5.1	5.3	0	Yes
500-hPa Relative Humidity (%)	80.3	79.7	0	Yes
850-hPa Relative Humidity (%)	67.4	66.4	0	Yes
Surface Dewpoint Temperature ($^{\circ}\text{C}$)	70.4	68.7	0	Yes
500-hPa Dewpoint Temperature ($^{\circ}\text{C}$)	80.3	79.7	0	Yes
850-hPa Dewpoint Temperature ($^{\circ}\text{C}$)	67.4	66.4	0	Yes

Table 4.9: Summary statistics of changes in cell properties in urban and downwind regions. ML = mixed layer, MU = most unstable layer. Both MU and ML are within the lowest 50 hPa.

Environmental metric	Urban (mean)	Downwind (mean)	p-value	Significant?
CAPE (J/kg)	2125	2170	3.57×10^{-4}	Yes
CIN (J/kg)	-34	-36	1.81×10^{-3}	Yes
ML Temperature ($^{\circ}\text{C}$)	27.5	27.7	1.33×10^{-14}	Yes
ML Dewpoint Temperature ($^{\circ}\text{C}$)	22.4	22.4	1.34×10^{-2}	Yes
ML CIN (J/kg)	-3	-4	4.86×10^{-3}	Yes
MU Temperature ($^{\circ}\text{C}$)	28.8	29.2	4.35×10^{-25}	Yes
MU Dewpoint Temperature ($^{\circ}\text{C}$)	23.3	23.3	0.172	No
MU CAPE (J/kg)	2205	2239	3.29×10^{-3}	Yes
MU CIN (J/kg)	-28	-32	9.95×10^{-9}	Yes
0-1km Storm Relative Helicity (m^2/s^2)	-1.4	0.1	3.12×10^{-11}	Yes
0-3km Storm Relative Helicity (m^2/s^2)	10.3	11.6	8.45×10^{-7}	Yes
0-6km Storm Relative Helicity (m^2/s^2)	21.2	19.7	1.02×10^{-9}	Yes
0-1km Bulk Shear (m/s)	4.3	4.2	2.31×10^{-5}	Yes
0-3km Bulk Shear (m/s)	5.1	5.4	1.42×10^{-12}	Yes
0-6km Bulk Shear (m/s)	5.9	5.6	8.61×10^{-8}	Yes
Surface Wind Speed (m/s)	4.7	5.1	1.05×10^{-37}	Yes
500-hPa Wind Speed (m/s)	3.0	3.0	0	Yes
850-hPa Wind Speed (m/s)	5.3	5.3	0	Yes
500-hPa Relative Humidity (%)	79.7	79.6	0	Yes
850-hPa Relative Humidity (%)	66.4	66.3	0	Yes
Surface Dewpoint Temperature ($^{\circ}\text{C}$)	68.7	68.5	0	Yes
500-hPa Dewpoint Temperature ($^{\circ}\text{C}$)	79.7	79.6	0	Yes
850-hPa Dewpoint Temperature ($^{\circ}\text{C}$)	66.4	66.3	0	Yes

4.5 Summary of Cell Properties in the Regions

The analysis of convective activity across the Houston area reveals distinct regional characteristics influenced by local processes and circulations. The upwind region, primarily driven by sea breeze circulation, exhibits smaller cell areas and moderate precipitation compared to other regions. In contrast, the urban region, influenced by the UHI effect, experiences more frequent and sustained convective initiation, larger cell areas, and variable precipitation patterns. The downwind region also shows larger cell areas and the highest precipitation efficiency among the three regions. Temporally, upwind convective initiation peaks during the night through early morning hours but decreases in the afternoon. The urban region exhibits a pronounced peak in initiation during late morning to early afternoon due to enhanced surface heating from the UHI, with sustained activity into the evening hours. The downwind region displays a complex, multimodal diurnal cycle, with the highest initiation occurring in the late evening and night.

Examining cell tracks transitioning across regions provides further insights. Cells moving from the upwind to the urban region did not reveal statistically significant property changes, suggesting the initial urban transition may not immediately impact cell structure and composition. However, cells propagating from the urban to downwind regions showed significant decreases in area, reflectivity, and echo top height, indicating the urban environment can influence the evolution and dissipation of cells as they move downwind.

4.5.1 Radar Sampling and Resolution Effects

While the regional comparisons provide valuable insights into urban influences on convective processes, it is important to consider the potential impact of radar sampling and

resolution effects on our observations. Echo top height measurements are particularly susceptible to these effects (Mandement et al., 2023). Echo tops are typically determined by identifying the highest elevation angle where reflectivity exceeds a certain threshold (in this case, 20 dBZ). However, this process is complicated by the radar’s scanning strategy, beam width, and distance from the target. In the urban region where the KHGX radar is located, we observe a reduction in 20-dBZ echo top heights. This reduction is likely an artifact of the radar’s scanning strategy rather than a true storm feature. The primary issue here is the ‘cone of silence’ effect, where the radar’s highest elevation angles may not capture the full vertical extent of nearby storms. For instance, at a distance of about 15 km from the radar, the maximum measurable echo top height is limited to approximately 6 km due to the radar’s scanning strategy. This limitation leads to a potential underestimation of echo top heights in the urban region. Conversely, in the upwind and downwind regions, which are farther from the radar, the radar beam samples a greater vertical extent of the atmosphere. This allows for more accurate measurements of echo top heights in these areas, as there are radar scans at almost every level of the storm.

The regridding process, especially in the vertical dimension, further impacts echo top estimates, as it significantly affects how data from different elevation angles is combined and interpolated. This process can be especially challenging near the radar where the ‘cone of silence’ exists. In the ‘cone of silence’, there is a lack of data above a certain height due to the radar’s maximum elevation angle. During regridding, this absence of data can lead to interpolation. For instance, the algorithm might interpolate between the highest available measurement and an assumed echo-free area above, potentially underestimating the true echo top height. The impact of regridding on echo top estimates varies with distance from the radar. Close to the radar, where the ‘cone of silence’ is most pronounced, regridding might fail to accurately represent

the full vertical extent of storms. Farther from the radar, where more elevation angles intersect the storm structure, regridding can provide a more accurate vertical profile, albeit with decreased horizontal resolution. Thus, while there may be some loss of resolution at farther ranges, the primary limitation for echo top height estimation appears to occur close to the radar. Future studies could further investigate the effects of regridding on echo top estimates.

Moreover, the KHGX NEXRAD radar used in this study operates with a 1° angular beamwidth, which introduces range-dependent variations in sampling volume. At close ranges, the radar has a high spatial resolution, potentially leading to more accurate depictions of cell areas and structures. However, as the range increases, the beam widens, resulting in larger sampling volumes (Ryzhkov, 2007). For instance, at a range of 50 km, the beam width is approximately 0.87 km, while at 150 km, it expands to about 2.62 km. This beam broadening effect can artificially inflate cell areas at farther ranges, as smaller convective cells may be merged within a single beam. Consequently, our observed cell areas may be overestimated for storms further from the radar, particularly in the upwind and downwind regions.

Reflectivity measurements also suffer from range-dependent biases. Beam filling issues at far ranges can lead to underestimation of maximum reflectivity values, as the peak reflectivity core may only partially fill the beam. Additionally, the minimum detectable signal increases with range due to power loss, potentially causing us to miss weaker echoes at the storm periphery for distant cells.

To mitigate these effects, we could implement a range-normalization procedure in future studies, adjusting cell properties based on their distance from the radar. Utilizing data from multiple nearby radars could also help to reduce range-dependent biases.

4.5.2 Gridding Effects on Storm Structure Representation

While Section 4.5.1 addressed the general impacts of radar sampling and resolution, we also need to consider how our specific gridding process affects the representation of storm structures in our data. Our study utilizes a 500-m Cartesian grid with the Barnes 2 objective analysis scheme, which has varying implications across the study area due to changing radar beam characteristics with range.

In the close range, primarily covering the urban region within 30 km of the radar, the 500-m grid spacing results in gridded data that closely represents the storm's true structure, especially in the lower levels. For instance, at 20 km range, the beam width is only about 350 m, allowing for detailed sampling of storm structures consistent with the grid spacing. At far ranges beyond 100 km, radar resolutions interact differently with our 500-m grid spacing. The radial resolution, at approximately 150 m, remains finer than the grid spacing, being about 0.3 times its size. In contrast, the azimuthal resolution expands to approximately 1.75 km, exceeding the grid spacing by a factor of about 3.5. This disparity between radial and azimuthal resolutions relative to our grid affects the representation of storm structures in our data. In these areas, the gridding process relies heavily on interpolation and extrapolation, potentially smoothing out small-scale features of the storms.

Beam altitude can also significantly impact our gridded data quality across these ranges. While low-level features are well-captured close to the radar, the situation changes dramatically with distance. For example, at 100 km range, even the lowest beam (0.5° elevation) samples at approximately 1.5 km above ground level, missing crucial low-level information. These factors have several implications for our analysis, complementing those discussed in Section 4.5.1. The vertical structure representation becomes increasingly range-dependent, with storms closer to the radar having better-resolved vertical structures, especially in the lower levels. Small-scale features are more

likely to be captured in near-range storms and smoothed out in far-range storms. Thus, the differences in cell initiation patterns between urban and downwind regions (Fig. 4.1) may be partially influenced by these gridding effects, in addition to the radar sampling effects discussed earlier and genuine meteorological factors.

While our gridding process improves overall data quality, it cannot fully compensate for the fundamental limitations of beam propagation and sampling geometry. Therefore, when comparing storm structures across different ranges, particularly between near and far ranges, we must keep in mind these gridding-induced variations in data representation.

Chapter 5

Microphysical Properties of the Convective Cells

5.1 Vertical Structure and Characteristics of the Cells

The normalized contoured frequency by altitude diagrams (CFADs) of the observed polarimetric variables provide insights into the microphysical processes occurring within the cells in the Houston area during the Summer of 2022 (Fig. 5.1). These CFADs are constructed by binning the polarimetric radar data into altitude and bins (see Table 3.4 in Chapter 3), and then normalizing the frequency by maximum count across all heights. To calculate the average CFADs for each region, we employed a straightforward method. For each polarimetric variable in a given region, we summed the CFAD values from all cells within that region and then divided by the total number of cells. By averaging the CFADs, we can identify consistent patterns and characteristics that are robust across multiple storm events, while smoothing out individual cell-to-cell variability.

In the analysis of these CFADs, the melting layer can be identified at approximately 4 km above ground level (AGL). This identification is based on the decrease in ρ_{hv} values, which exhibits a slight bulge to the left at this height (Kumjian, 2013b). The melting layer is a significant feature as it marks the transition zone where ice particles

melt into raindrops, thus affecting the polarimetric signatures and microphysical processes within the storm. While the bright band is typically expected to be a maximum in reflectivity in stratiform rain regions, the CFADs in this case show similar reflectivity values below the melting layer.¹ This observation suggests that the melting process within these cells may not be as pronounced or as uniform as in stratiform precipitation systems. Possible explanations for this behavior could include the presence of strong updrafts that disrupt the accumulation of melting hydrometeors or the rapid melting of smaller ice particles, leading to a less pronounced bright band signature.

In Fig. 5.1(a), the average Z exhibits a wider distribution near the surface than aloft, with values decreasing with height. This is consistent with the expected microphysical processes in convective storms, where ice particles nucleate and grow in the upper regions and raindrops form through melting and grow through coalescence processes in the lower levels. Notably, Z exhibits a unimodal distribution from the surface to the top of the storm, and the peak in the frequency distribution below the melting layer ($\sim 45\%$) occurs around reflectivity values of 20–40 dBZ. Meanwhile, this peak is also observed above the melting layer up to 10 km but at a relatively smaller but wider range of Z values between 10 and 35 dBZ. When examined level by level, the average Z distribution above the melting layer progressively shifts towards lower values at higher altitudes. These lower Z values above the melting layer compared to those below indicate that the ice particles aloft backscatter less power compared to the liquid hydrometeors (raindrops) below the melting layer (Kumjian, 2013a).

¹It is well known that convective regions do not typically have clearly defined bright band signatures.

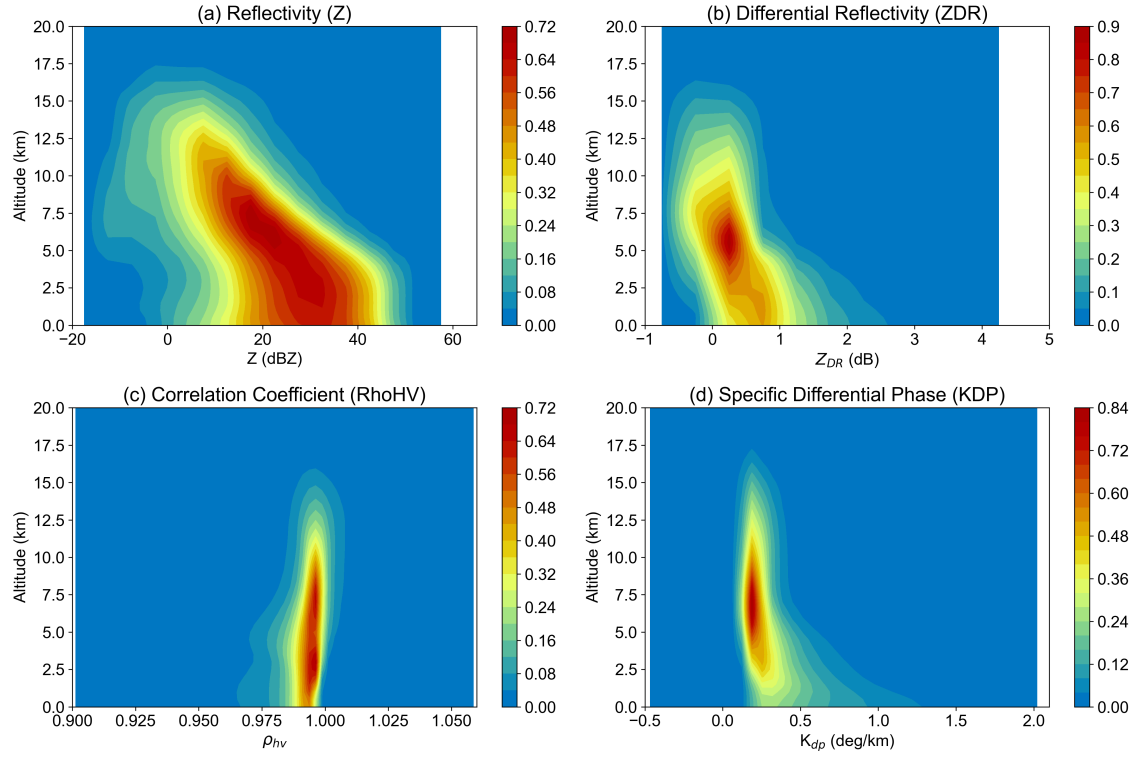


Figure 5.1: Normalized average CFADs of polarimetric variables for the entire region. (a) Horizontal reflectivity (Z), (b) differential reflectivity (Z_{DR}), (c) co-polar correlation ratio (ρ_{hv}), and (d) specific differential phase (K_{DP}). The color bar indicates the frequency.

Meanwhile, Z_{DR} also reveals a unimodal distribution that further narrows and shifts to the left above the melting layer, indicating different shapes and orientations of hydrometeors (Fig. 5.1(b)). Although there are some hydrometeors with negative Z_{DR} above the melting layer, most of the particles above the melting layer have positive Z_{DR} values, with the peak frequency (70%) centered around 0.4 dB between 4.5 km and 7.5 km AGL. This high frequency of positive Z_{DR} values implies that the majority of the hydrometeors just above the melting layer are small, oblate (horizontally oriented), and have a consistent shape, size, and orientation. However, below the melting layer, the

average Z_{DR} distribution broadens and shifts to higher values, peaking at around 2.5 dB. This increase in Z_{DR} values below the melting layer indicates the presence of large, oblate hydrometeors, typically raindrops, which are typically associated with heavy rain (Kumjian, 2013a). The broadening of the Z_{DR} distribution below the melting layer also suggests a greater variability in raindrop sizes and shapes in the lower levels of the storm. This variability is likely due to the coalescence and breakup processes that occur during raindrop formation and growth.

The average ρ_{hv} exhibits a relatively narrow distribution throughout the vertical extent of the storm compared to Z and Z_{DR} , with most of the values between 0.98 and 1.00 (Fig. 5.1(c)). This high value implies uniformity in the physical characteristics of the scatterers, confirming that they are hydrometeors rather than non-meteorological scatterers, which typically have ρ_{hv} values less than 0.8 (Kumjian, 2013a). The narrow distribution above the melting layer suggests a consistent composition of hydrometeors aloft in this region, resulting in a high degree of correlation between the radar signals in the horizontal and vertical polarizations. However, below the melting layer, the ρ_{hv} distribution slightly broadens, indicating a greater variability in the shape and composition of the hydrometeors in the lower levels of the storm. This broadening is primarily attributed to the melting of ice particles into raindrops. This melting process can lead to irregularities in the raindrop shape, causing deviations from the ideal oblate spheroidal shape assumed for liquid hydrometeors.

The average K_{DP} also exhibits a unimodal distribution throughout the vertical extent of the storm, with a peak value of around 1.5 deg/km near the surface (Fig. 5.1(d)). There is a high frequency around 0.1-0.2 deg/km concentrated between 5 km and 10 km AGL. While K_{DP} is less sensitive to ice crystals compared to raindrops, it can still detect the presence of asymmetric ice particles, especially in the presence of strong electric fields (Kumjian, 2013a). Several studies have shown that K_{DP} can be

used to detect the alignment of small ice crystals in the horizontal or vertical direction, producing positive or negative K_{DP} values aloft (Hubbert et al., 2014; Wang et al., 2024). Notably, the K_{DP} distribution broadens from 7 km AGL downwards, consistent with the findings of (Shrestha et al., 2022). Here, it is likely that ice particles are beginning to melt, with some melting at a higher rate than others, resulting in variability in the rainfall rates. The variability can also be due to ongoing microphysical processes, such as coalescence and breakup, that influence the size and shape of raindrops as they grow and evolve within the storm. The increase in K_{DP} values below the melting layer is consistent with high rainfall rates and the presence of heavy precipitation observed in the Z plot (Fig. 5.1(a)).

The CFADs presented so far provide valuable insights into the microphysical processes occurring within the convective cells observed over the Houston area during the Summer of 2022. However, it is essential to examine these polarimetric signatures across different regions to understand the potential impact of different land environments on the microphysical processes and storm evolution.

5.2 CFAD Analysis by Region

5.2.1 CFAD Analysis in the Upwind Region

The normalized average CFADs of Z , Z_{DR} , ρ_{hv} , and K_{DP} in the upwind region are shown in Figs. 5.2(a)-5.5(a). Below the melting layer, the majority of the Z CFAD values are concentrated between 20-40 dBZ, while most values above the melting layer fall within the range of 10-40 dBZ (Fig. 5.2(a)). Notably, above 15 km AGL, the frequencies ($\sim 25\%$) are significantly lower compared to lower altitudes ($> 60\%$), indicating that very few storms extend beyond this height. Therefore, we consider 15 km as the storm

top in this region to avoid potential errors in characterizing the properties above this altitude due to the small sample size.

The average Z CFAD shows high frequencies of hydrometeors at all levels of the storm, contrasting with the downwind region where the distribution varies at different heights. Above the melting layer (e.g., ~ 7.5 km AGL), the peak in the average Z in this region shows a narrower spread between 5 and 25 dBZ compared to the urban region ($Z \sim 0-25$ dBZ) and a similar spread to that in the downwind region. Additionally, the average Z_{DR} has a similar spread in this region as in other regions above the melting layer (Fig. 5.3(a)). These observations suggest that the hydrometeors aloft in this region tend to have less variability in number concentration (due to the lower spread in average Z values) than in the urban region, with the shapes and orientations of these hydrometeors being similar across all regions due to the similar average Z_{DR} values.

Moreover, there is a high frequency of K_{DP} values less than 0.2 deg/km extending from 4 km AGL to about 10 km AGL, indicating high concentrations of hydrometeors with low ice water content above the melting layer (Fig. 5.5(a)). Below the melting layer and near the surface, the spread in average Z in this region ($Z \sim 10-50$ dBZ) is lower than in the urban region ($Z \sim 0-50$ dBZ), with a higher frequency of average $Z_{DR} < 1$ dB in the upwind region compared to other regions (Fig. 5.3(a)). However, the Z_{DR} spread is not as wide as in other regions, with the maximum Z_{DR} being 2.5 dB in this region compared to 4 dB and 3.5 dB in the urban and downwind regions, respectively. The slightly higher frequencies at $Z_{DR} < 1$ dB below the melting layer in this region than in other regions may suggest that the upwind region likely has a slightly greater occurrence of small raindrops and fewer large drops compared to other regions. Therefore, the upwind region tends to have a more uniform microphysical environment, where small drops are more likely to occur than in other regions. This

is consistent with the expected behavior of convective storms in the absence of very strong external influences (see Section 4.4.3 in Chapter 4).

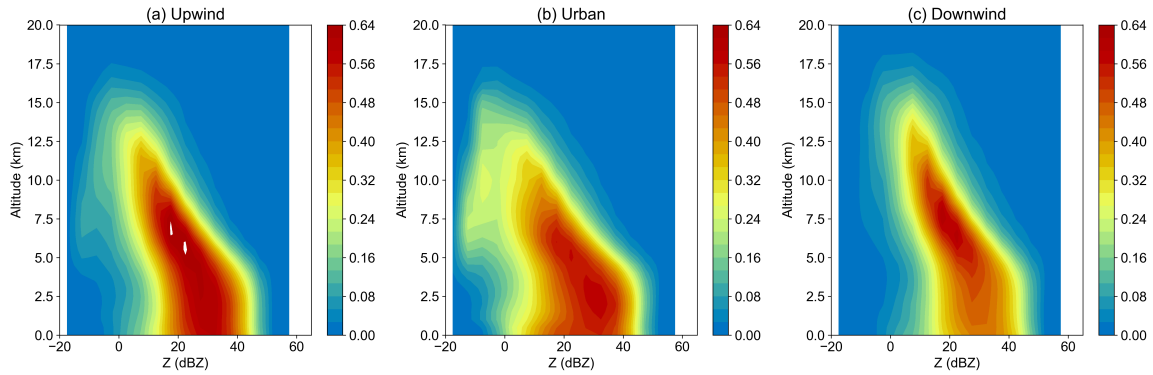


Figure 5.2: Normalized averaged CFADs for the horizontal reflectivity based on region

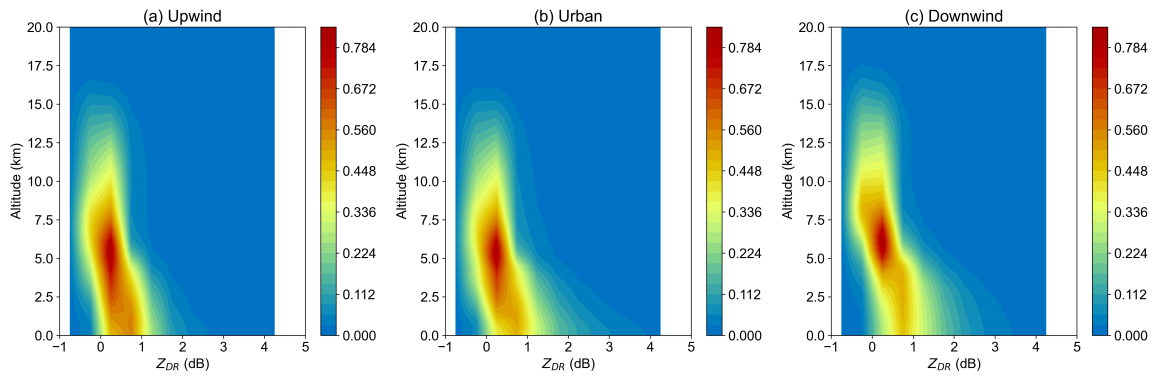


Figure 5.3: Normalized averaged CFADs for the differential reflectivity based on region

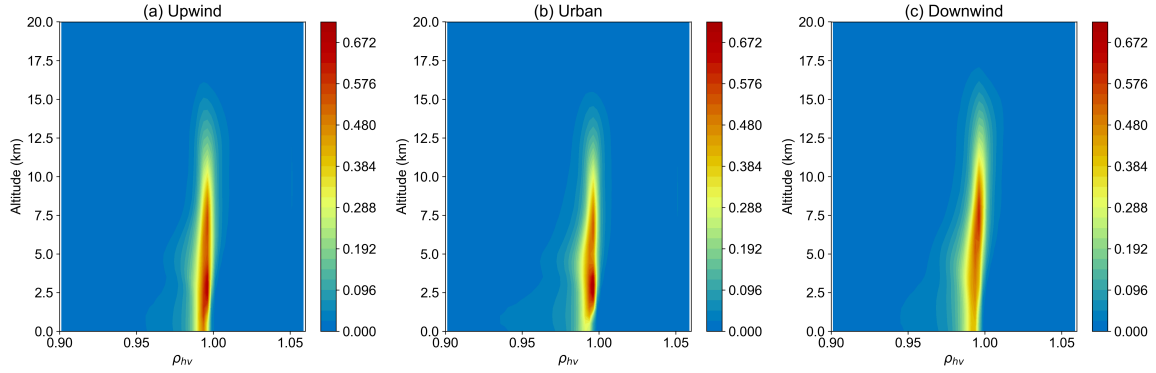


Figure 5.4: Normalized averaged CFADs for the co-polar correlation coefficient based on region

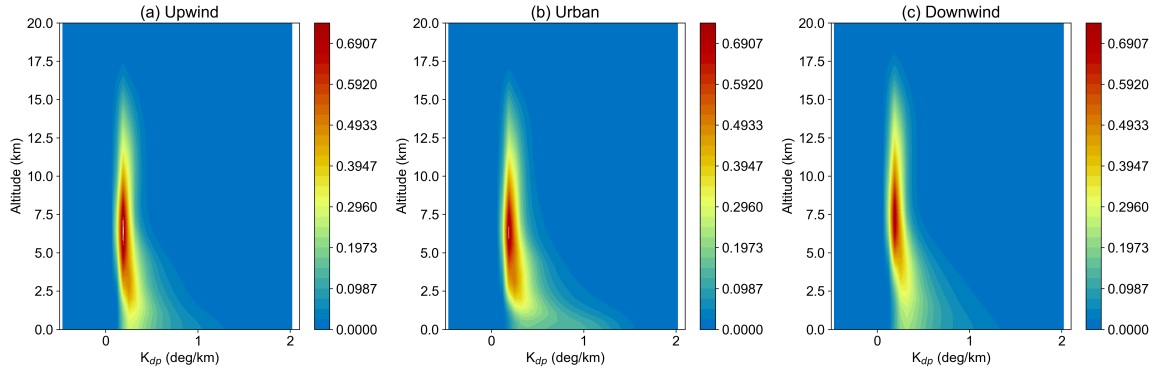


Figure 5.5: Normalized averaged CFADs for the specific differential phase based on region

5.2.2 CFAD Analysis in the Urban Region

In the urban region, the normalized average CFADs of Z , Z_{DR} , ρ_{hv} , and K_{DP} are shown in Figs. 5.2(b)-5.5(b). Similar to the upwind region, the storm top in the urban region is approximately 15 km AGL owing to the low frequencies above this height. The average Z CFAD exhibits a wider spread compared to the upwind and downwind regions throughout the vertical structure of the storm, suggesting greater variability in

the size and concentration of hydrometeors within the urban environment (Fig. 5.2(b)). Below the melting layer, a wider spread of average Z values between 18 and 40 dBZ is observed in this region compared to other regions. Moreover, there is a broader average Z_{DR} distribution of 0-4 dB below the melting layer in this region compared to 0-2.5 dB in the downwind region (Fig. 5.3(b)). These observations indicate higher variability in the concentration (due to the wide spread in Z), shape, and size (due to the wider spread in Z_{DR}) of hydrometeors in the urban region than other regions. In fact, raindrops in this region tend to have the largest size (maximum $Z_{DR} \sim 4$ dB) compared to other regions (maximum $Z_{DR} < 4$ dB), and they are associated with increased variability in rainfall rates, as indicated by the wider K_{DP} distribution (Fig. 5.5(b)). Owing to the wider spread in Z and Z_{DR} in this region at all levels of the storm, the urban region is likely characterized by hydrometeors with more heterogeneous shapes and sizes compared to other regions throughout the storm depth.

5.2.3 CFAD Analysis in the Downwind Region

In the downwind region, the normalized average CFADs of Z , Z_{DR} , ρ_{hv} , and K_{DP} are shown in Figs. 5.2(c)-5.5(c). Similar to the upwind region, the frequency distribution of the average Z and Z_{DR} values is narrower throughout the storm depth in this region compared to the urban region (Fig. 5.2(a)). Notably, the spread of average Z values near the surface in the downwind region is 18-50 dBZ compared to the 10-50 dBZ in the urban region. This suggests more uniform microphysical characteristics and less variability in the size (because Z_{DR} spread is lower) and concentration (due to lower spread in Z) of hydrometeors in this region compared to the urban region. However, a notable difference from the upwind region is that the downwind region has different distributions of average Z values above and below the melting layer, whereas the upwind region has similar frequency distributions at all levels of the storm. Meanwhile, Z_{DR}

values exhibit a maximum value of 3.5 dB compared to the 2.5 dB near the surface, suggesting that although the difference in the sizes of hydrometeors in both regions is small, the downwind region may have slightly larger raindrops near the surface than the upwind region. Above the melting layer, the average Z_{DR} is similar across all regions, suggesting similar shapes, sizes, and orientations of hydrometeors aloft across all the regions.

5.3 Comparing Microphysical Processes across Regions

As precipitation particles descend towards the Earth’s surface, they undergo various microphysical processes that can alter their sizes, spatial distribution, and phase. Polarimetric variables are capable of detecting these changes due to their sensitivity to particle sizes, shapes, number concentrations, and physical compositions. Since these modifications primarily occur as hydrometeors fall outside of strong updrafts, analyzing the vertical profiles of polarimetric radar variables can provide valuable insights into the ongoing microphysical processes (Kumjian et al., 2022).

For this part of our analysis, we consider the storm top to be at 15 km AGL, as the frequency of storms in the CFADs decreases significantly above this level in all the regions (Fig. 5.2). To better understand the microphysical processes occurring in each region, we will examine the rain and ice microphysical processes separately.

5.3.1 Microphysical Processes in the Upwind Region

5.3.1.1 Rain microphysical processes

The vertical mean profiles of the polarimetric variables in the upwind region are shown in Fig. 5.6. Below the melting layer, the upwind region experiences an increase in average Z_{DR} ($Z_{DR} \sim 0.85$ dB), indicating the presence of collision-coalescence processes. During this process, two or more particles collide and merge to form larger drops, resulting in an increase in drop size (Kumjian et al., 2022). Traditionally, both Z_{DR} and Z profiles serve as indicators for collision-coalescence occurrence; an increase in Z with decreasing height is often associated with such processes, as larger drops formed through coalescence contribute more to reflectivity. However, in this case, while we observe an increase in Z_{DR} , the average Z remains relatively constant ($Z \sim 28$ dBZ) from the melting layer to the surface. This lack of a clear increasing trend in the Z profile suggests that the collision-coalescence process, if present, may not be as pronounced or widespread as typically observed in strong coalescence cases.

An increased K_{DP} ($K_{DP} \sim 0.6$ deg/km) near the surface, revealing increased rainfall rates. However, this effect is not as pronounced as in the urban region, which exhibited a relatively higher Z_{DR} ($Z_{DR} \sim 1.2$ dB) and K_{DP} ($K_{DP} \sim 0.95$ deg/km) near the surface (Fig. 5.6(b,d)). While the average Z is slightly higher ($Z \sim 28$ dBZ) than that in the urban region ($Z \sim 25$ dBZ), it remained consistent from the melting layer to the surface. This may suggest that the concentration of hydrometeors remained relatively consistent below the melting layer in the upwind region. Furthermore, a relatively constant average ρ_{hv} value of ~ 0.98 suggests that pure rain typically occur below the melting layer and near the surface (Kumjian, 2013a).

Overall, while other factors may also be contributing to the observed patterns, collision-coalescence is likely occurring below the melting layer in the upwind region,

as indicated by the increasing values of average Z_{DR} and K_{DP} toward the surface. Despite rain drops coalescing into large drops in this region, these drops do not tend to increase in size as substantially as those in other regions, particularly the urban region.

5.3.1.2 Ice Microphysical Processes

From the storm top to approximately 7 km AGL, there is an increase in average Z ($\Delta Z \sim 13$ dBZ) toward the ground, with K_{DP} remaining constant at 0.1 deg/km (Fig. 5.6(a),(d)). Meanwhile, Z_{DR} slightly increases by 0.1 dB before decreasing from 0.2 to 0.15 dB. Several microphysical processes are likely occurring between these height levels. First, it is speculated that the increase in average Z near the storm top may indicate vapor depositional nucleation and growth, with ice particles gaining mass from the surrounding vapor (Kumjian et al., 2022). The positive but slight increase in Z_{DR} likely indicates that the ice particles are aligned in the horizontal dimension and are very small, given the small Z_{DR} value of 0.2 dB at 14 km AGL. Second, riming and aggregation processes may also be occurring along with vapor depositional growth, given the increase in Z and decrease in Z_{DR} . In the former case, the ice crystals may continue to collect and accumulate cloud droplets that freeze upon contact, leading to higher average Z . In the latter case, two or more particles may coagulate to form a larger particle, causing the average Z to increase. It should be noted that other microphysical processes apart from those stated here may also come into play, but these processes discussed above are clearly evident above the melting layer.

Meanwhile, there is a steady increase in Z_{DR} and K_{DP} from 7 km AGL down to the melting layer, which may also suggest the presence of a Z_{DR} or K_{DP} column (Kumjian, 2013b). These columns refer to areas of enhanced Z_{DR} and K_{DP} values that extend above the melting layer. The presence of Z_{DR} and K_{DP} columns indicates large supercooled drops and wet ice particles being lifted by strong updrafts to subfreezing

temperatures (Kumjian, 2013b). These droplets may remain in a liquid state for a period before freezing, resulting in the observed Z_{DR} and K_{DP} enhancements above the melting layer in the updraft region.

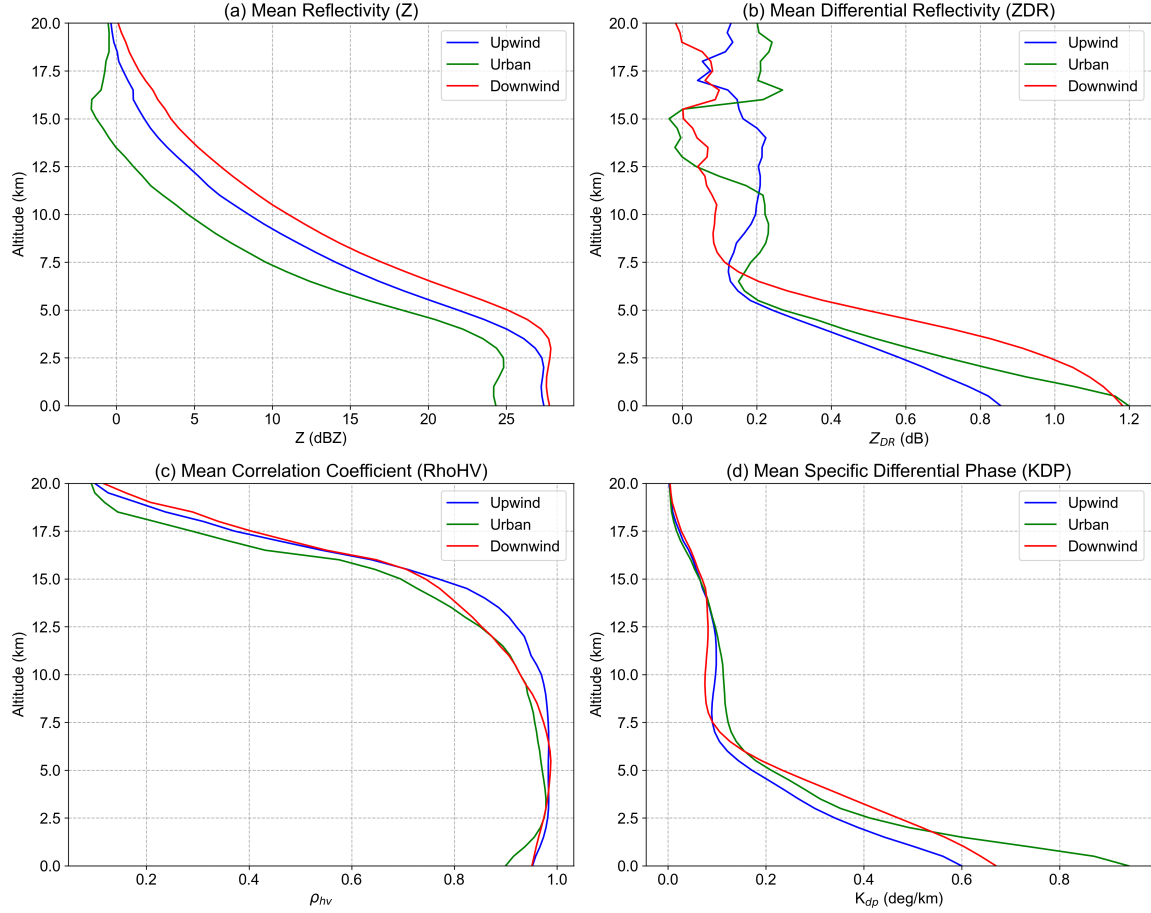


Figure 5.6: Vertical mean profiles of polarimetric variables based on region. (a) horizontal reflectivity (Z), (b) differential reflectivity (Z_{DR}), (c) co-polar correlation coefficient (ρ_{hv}), and (d) specific differential phase (K_{DP})

5.3.2 Microphysical Processes in the Urban Region

5.3.2.1 Rain microphysical processes

The urban region exhibits increases in average Z_{DR} and K_{DP} below the melting layer, with both values near the ground being almost 50% higher than those in the upwind region (Fig. 5.6(b),(d)). Specifically, the Z_{DR} and K_{DP} had average values of 1.2 dB and 0.9 deg/km, respectively. Despite the constant Z of approximately 25 dBZ below the melting layer, the enhanced Z_{DR} and K_{DP} values suggest that the urban region generally tends to have more pronounced collision-coalescence processes than the upwind region. Notably, the ρ_{hv} value is about 0.9 (Fig. 5.6(c)), which is less than the 0.98 threshold for pure rain. Hence, mixed-phase hydrometeors, such as wet hail or graupel, may occur in this region, in addition to the raindrops below the melting layer. The highest K_{DP} (0.9 deg/km) in this region compared to other regions suggests that the highest rainfall rates occur in this region.

Overall, in comparison to the upwind region, the urban area demonstrates evidence of more efficient collision-coalescence processes below the melting layer and near the ground. Furthermore, the data indicates a tendency for mixed-phase precipitation to occur in this region, contributing to its distinct precipitation characteristics.

5.3.2.2 Ice Microphysical Processes

As in the upwind region, various ice microphysical processes occur from the storm top at 15 km AGL to ~ 7 km AGL. First, the increase in average Z ($\Delta Z \sim 15$ dBZ) and slight increase in Z_{DR} ($\Delta Z_{DR} \sim 0.2$ dB) likely indicate nucleation and initial growth by vapor deposition near the storm top. This is supported by the low ρ_{hv} values of 0.8–0.9 between 15 km and 11 km AGL, which suggest that the nucleated particles may be wet hail or ice crystals (Kumjian, 2013a). However, the average Z_{DR} and K_{DP}

(Fig. 5.6(d)) remain relatively constant from 11 km to 7 km AGL, showing evidence of riming. ρ_{hv} also increases to ~ 0.95 at 7 km AGL, which may indicate the presence of graupel at this height level. Additionally, the increase in Z suggests that aggregation might also be occurring alongside the riming process. Meanwhile, the increases in Z_{DR} and K_{DP} from 7 km AGL to the melting layer could suggest the presence of a Z_{DR} or K_{DP} column, similar to the upwind region (Kumjian, 2013a).

Overall, different processes such as vapor deposition, riming, and aggregation are evident in this region. While the polarimetric variables exhibit some differences in values compared to those in the upwind region, these differences are quite small, and there appear to be similar ice microphysical processes between the upwind and urban regions.

5.3.3 Microphysical Processes in the Downwind Region

5.3.3.1 Rain Microphysical Processes

In the downwind region, the average Z and ρ_{hv} values near the surface are similar to those in the upwind region, but the Z_{DR} value is substantially higher. Specifically, the average Z_{DR} value is 1.2 dB near the ground, approximately 35% higher than that in the upwind region. This increase in Z_{DR} suggests a more efficient collision-coalescence process in the downwind region compared to the upwind region, resulting in larger average rain drop sizes. Notably, these drop sizes are comparable to those observed in the urban region (Fig. 5.6(b)). However, similar to the upwind region, the average Z remains relatively constant ($Z \approx 28$ dBZ) from the melting layer to the surface.

Meanwhile, the lower average K_{DP} in the downwind region compared to the urban region may indicate less liquid water content and lower rainfall rates. A possible explanation for the differences in K_{DP} values between these regions is that the urban region

may have more mixed-phase hydrometeors, which produce larger rainfall amounts when they melt. Nevertheless, it is important to note that other factors can also influence the observed lower rainfall rates in the downwind region.

5.3.3.2 Ice Microphysical Processes

From the storm top at 15 km AGL to ~ 8 km AGL, there is a substantial increase in average Z from 5 dBZ to 15 dBZ, with the average Z_{DR} and K_{DP} values increasing very slightly. This suggests that ice particles are likely growing primarily by vapor deposition, similar to other regions. However, the small changes in average Z_{DR} and K_{DP} at these levels may also indicate contributions from riming and aggregation processes to the ice microphysics. The increase in average Z without significant changes in Z_{DR} and K_{DP} could be attributed to an increase in the number of ice particles rather than substantial changes in their shape or orientation. This is consistent with the understanding that numerous ice particles can lead to high Z values, especially if the particles are relatively large and dense.

Meanwhile, the average ρ_{hv} increased from 0.8 at the storm top to ~ 0.95 at 8 km AGL, exhibiting higher values than in other regions at these levels. This high ρ_{hv} might indicate the presence of dry hail or dry graupel, as these particles often exhibit very high ρ_{hv} values (Kumjian, 2013a). The presence of these particles in large amounts could lead to collision processes that generate ice splinters, maintaining high ρ_{hv} values. This phenomenon, known as secondary ice production, can significantly influence the microphysics of the storm. Additionally, turbulence in this region of the storm could be a key driver of ice aggregation and riming processes, further contributing to the observed radar signatures.

Chapter 6

Summary and Conclusions

6.1 Summary of Research Findings

This study examined the characteristics and evolution of convective cells across the Houston metropolitan area, revealing significant urban and coastal influences on convective processes. Houston was chosen for this study due to its unique combination of a large urban area with strong heat island effects and its proximity to the Gulf of Mexico, which generates sea breeze circulations. This coastal-urban environment provides an ideal setting to investigate how the interaction between urban surfaces and coastal processes impacts convective storm evolution. The research aimed to investigate how different environments modify the development, evolution, and characteristics of convective cells. In this study, Houston was divided into three regions: upwind from the urban area, downwind from the urban area, and urban area. The analysis of 22,372 cell tracks across upwind, urban, and downwind regions provided comprehensive insights into the spatial and temporal dynamics of cells in various regions, as well as storms transitioning into and out of the urban area.

6.1.1 Upwind Region

The upwind region serves as the initial interaction zone between maritime air masses and the approaching urban landscape, and it is significantly influenced by offshore processes, particularly the sea breeze circulation. In this region, cell initiation peaks during the early morning hours, with approximately 80% of the total cells forming during this period. This frequency steadily reduces to 25% through the afternoons before increasing again into the night. The diurnal cycle of the mean cell area also shows variability, peaking at about 33 km^2 around 2 AM, then decreasing to approximately 23 km^2 by 5 AM. While the sea breeze is the primary driver of cell initiation in the early morning, moderate increases are observed as diurnal heating progresses. Meanwhile, cells in this region are generally smaller (42.5 km^2) and less intense (36 dBZ) compared to those in other regions.

The upwind region displays less microphysical variability with distinct features compared to urban and downwind regions. Normalized average CFADs and vertical profiles of Z , Z_{DR} , ρ_{hv} , and K_{DP} reveal a predominance of smaller raindrops in this region and lower overall rainfall amounts compared to other regions. These microphysical characteristics are evident throughout the cell lifecycle, from initiation to dissipation. During the developing stage, upwind cells exhibit strong initial convection, with reflectivity values reaching nearly 30 dBZ at the surface. However, as cells mature and begin to dissipate, they maintain smaller and more spherical raindrop sizes ($Z_{DR} \sim 0.9 \text{ dB}$) and lower rainfall rates ($K_{DP} \sim 0.65 \text{ deg/km}$) compared to cells in other regions (smaller drops being consistent with more tropical drop-size distributions). Collision-coalescence processes may occur below the melting layer (2.5 km AGL), but they are less pronounced in this region compared to urban and downwind areas, which have larger increases in Z_{DR} .

6.1.2 Urban Region

The urban region has a significantly higher number of initiated cells compared to other regions, primarily attributed to the impact of UHI on local atmospheric conditions. The number of cells per land area is approximately double compared to the upwind region and 4–5 times greater than the downwind region. In contrast to the upwind region, 30–40% of cells in this region are typically initiated during the period of peak heating (10 AM to 4 PM) and inward progression of the sea breeze, with sustained activity into the evening hours. While more storms initiated over the urban area, less precipitation per cell occurred. Thus, the total precipitation in Houston was not found to be significantly larger than other regions. Furthermore, there are numerous initiated cells that are less intense individually but with significant growth. Despite the high frequency of cell initiation, there is also a notable concentration of cell dissipation events within or near the urban environment. This suggests that many cells may not propagate far from their point of origin, possibly due to interactions with urban structures, localized downdrafts, or other urban-induced atmospheric perturbations. Overall, the cell lifetime in this region tends to be shorter (< 4 hours) compared to other regions (> 4 hours).

Microphysically, the average Z CFAD shows a broader distribution below the melting layer compared to other regions, indicating greater variability in hydrometeor sizes and concentrations. This variability is reflected in the cell lifecycle, where urban convective cells start with the lowest reflectivity ($Z \sim 27$ dBZ) in the developing stage but evolve to show the most intense precipitation in mature and dissipating stages (Z increasing to 47 dBZ). Unlike the upwind region, which has more numerous small, spherical raindrops, the urban region is characterized by a higher frequency of larger,

oblate raindrops associated with higher rainfall rates and concentrations. This is consistent with the observation that throughout its lifecycle, the urban region consistently exhibits higher liquid water content ($K_{DP} \sim 1.4$ deg/km) and larger raindrops ($Z_{DR} \sim 1.8$ dB) than other regions, especially during the mature stage. Meanwhile, the collision-coalescence process is more effective in the urban region than in the upwind region, leading to small particles merging into considerably large drops.

6.1.3 Downwind Region

The downwind region shows a lower number of cell dissipation events compared to the urban region, suggesting that while some cells may terminate within the urban area, others persist and potentially merge into fewer but larger cells as they move downwind. Precipitation efficiency is also a defining feature of the downwind region, with the highest total precipitation per cell (30 mm) among the three regions, indicating greater rainfall production efficiency. Despite a lower cell density (2 cells per 10 km² land area), the cells in this region tend to be more organized and intense, capable of producing substantial rainfall amounts. Meanwhile, the diurnal cycle in the downwind region is very complex, with multiple peaks in cell area observed throughout the day and night. Unlike the upwind and urban regions, less than 20% of the total cells are initiated in this region during the morning hours. However, this percentage increases significantly in the late afternoon to night (18–24 hours) to 40%, indicating that the majority of cells in this region are initiated during the late evening and nighttime hours.

The downwind region has the lowest frequency of average Z and Z_{DR} values near the surface among all regions, indicating that small, spherical raindrops are not as widespread in this region as in other regions. However, as cells mature, they exhibit the highest reflectivity ($Z \sim 47$ dBZ) among all regions and stages, indicating higher

precipitation intensity. This evolution aligns with the region’s high precipitation efficiency and organized, intense cell structure. Throughout the cell lifecycle, collision-coalescence process is particularly pronounced during the mature stage. Notably, the downwind region shows evidence of aggregation and riming below 7.5 km AGL even during the dissipating stage.

6.2 Contribution to Existing Knowledge

This study makes several significant contributions to the existing body of knowledge of urban impacts on convective processes. First, it advances our understanding of urban impacts on convection by identifying and characterizing distinct microphysical signatures in urban and downwind regions. The research reveals that urban areas have a broader distribution of raindrop sizes. These urban-induced microphysical modifications persist as storms propagate downwind, with the downwind region showing a narrower distribution of larger raindrops and extended vertical development of ice particles. These findings provide crucial insights into how urban environments alter precipitation formation mechanisms, contributing to the refinement of numerical weather prediction models for urban areas.

Second, this study demonstrates that urban influences on convective cells vary throughout their lifecycle. Enhanced convective initiation and early-stage intensity are observed in urban areas, followed by sustained higher intensities and larger sizes of urban-influenced cells in mature stages, and prolonged lifespans and intensity in dissipating stages. This lifecycle-based analysis provides a framework for understanding urban-convection interactions and offers a more nuanced perspective on how cities influence storm evolution. Moreover, the study improves understanding of how the

combined effects of the sea breeze and urban heat island influence the spatiotemporal evolution of initiated storms.

6.3 Limitations and Future Work

The study provides valuable insights into urban-convection interactions in Houston, but it is important to acknowledge its limitations and consider potential avenues for future research. While Houston offers an excellent case study, the generalizability of the findings is limited due to the city’s unique characteristics. Houston’s specific environmental conditions, including its humid subtropical climate, proximity to the Gulf of Mexico, and urban layout, may produce effects that are not representative of other urban environments (Caicedo et al., 2019; Zhang et al., 2018). These factors could influence convection patterns in ways that may differ from other metropolitan areas. Future research could extend this analysis to multiple urban areas with varying sizes and climates. A multi-city approach could lead to a more robust classification of urban impacts on convection based on city characteristics, including cities along and away from the coast. This expanded scope would help reveal which aspects of urban-convection interactions are generalizable and which are specific to certain urban types.

Furthermore, it is also important to consider longer-term effects. The current study is limited to a single summer season (June–August, 2022) due to computational and storage constraints, and it may not capture the full range of variability in convective patterns and urban influences over long timeframes. Weather patterns can vary significantly from year to year due to large-scale climate oscillations or long-term trends, which can potentially affect the urban-convection interactions observed in this study. To address this temporal limitation and complement the multi-city approach suggested

earlier, future work should involve multi-year studies examining seasonal and inter-annual variations in urban impacts on convective cells. Analyzing radar data over a period of 5–10 years or more could reveal how urban influences on convection may be modulated by large-scale climate patterns or how they might be changing over time due to urban growth or climate change. For example, NEXRAD data are available since the 1990s, providing an opportunity to evaluate how storm initiation or characteristics has changed as urban areas have expanded.

An important limitation of this study is the lack of direct aerosol measurements and their potential impact on convective processes in the urban environment and beyond. While the polarimetric radar data provided insights into microphysical processes, it could not directly quantify the influence of urban aerosols on cloud formation and precipitation. The observed differences in drop size distributions and ice particle characteristics across the upwind, urban, and downwind regions may be partially attributed to variations in aerosol concentrations, but this study was unable to isolate these effects. Future work could greatly benefit from leveraging data from recent field campaigns conducted in the Houston area, specifically ESCAPE and TRACER (Sheesley et al., 2023). These campaigns provide a wealth of aircraft and ground-based observations, including detailed aerosol measurements. While beyond the scope of this thesis, incorporating data from ESCAPE and TRACER could offer valuable insights into the role of aerosols in modulating convective processes in Houston’s urban and coastal environment. Furthermore, the potential impact of urban-generated aerosols on convective processes in the downwind region was not explicitly examined in this study. Urban aerosols can be transported by prevailing winds, potentially influencing cloud microphysics and precipitation processes well beyond the city limits. This transport of aerosols may explain some of the observed differences in cell characteristics

and evolution in the downwind region, but the current study lacks the data to confirm this hypothesis. Future work should incorporate aerosol measurements, such as particulate matter concentrations and aerosol optical depth, to better understand how urban-generated aerosols modify cloud microphysics and precipitation processes both within and outside the urban area. The ESCAPE and TRACER datasets could prove invaluable for this purpose. Additionally, future studies could explore the use of dual-polarization radar retrievals of raindrop size distributions in combination with in-situ aerosol measurements to investigate potential aerosol impacts on warm rain processes in the urban environment and downwind regions. Such research could help elucidate the relationships between urban aerosols, convective dynamics, and precipitation efficiency, potentially explaining some of the observed differences in rainfall patterns across the Houston metropolitan area and its surroundings.

Finally, the primary reliance on KHGX NEXRAD radar data, while providing excellent coverage of the Houston area, can be subject to beam blockage, especially near the ground, potentially biasing the findings regarding the microphysical processes. Future studies could benefit from using multiple radars or networks of high-resolution X-band radars to provide more comprehensive coverage and potentially reveal finer-scale features of urban-influenced convection.

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