Examining Polarimetric Radar Observations of Bulk Microphysical Structures and Their Relation to Vortex Kinematics in Hurricane Arthur (2014)

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ABSTRACT

Dual-polarization radar observations were taken of Hurricane Arthur prior to and during landfall, providing needed insight into the microphysics of tropical cyclone precipitation. A total of 30 h of data were composited and analyzed by annuli capturing storm features (eyewall, inner rainbands, and outer rainbands) and by azimuth relative to the deep-layer environmental wind shear vector. Polarimetric radar variables displayed distinct signatures indicating a transition from convective to stratiform precipitation in the downshear-right to downshear-left quadrants, which is an organization consistent with the expected kinematic asymmetry of a sheared tropical cyclone. In the downshear-right quadrant, vertical profiles of differential reflectivity $Z_{DR}$ and copolar correlation coefficient $\rho_{HV}$ were more vertically stretched within and above the melting layer at all annuli, which is attributed to convective processes. An analysis of specific differential phase $K_{DP}$ indicated that nonspherical ice particles had an increased presence in two layers: just above the melting level and near 8-km altitude. Here, convective updrafts generated ice particles in the lower layer, which were likely columnar crystals, and increased the available moisture in the upper layer, leading to increased planar crystal growth. A sharp transition in hydrometeor population occurred downwind in the downshear-left quadrant where $Z_{DR}$ and $\rho_{HV}$ profiles were more peaked within the melting layer. Above the melting layer, these signatures indicated reduced ice column counts and shape diversity owing to aggregation in a predominantly stratiform regime. The rainband quadrants exhibited a sharper transition compared to the eyewall quadrants owing to weaker winds and longer distances that decreased azimuthal mixing of ice hydrometeors.

1. Introduction

Improvements in forecasts of tropical cyclone (TC) intensity and structure require a better understanding of the microphysical processes occurring in the TC clouds and precipitation. Numerical models utilizing complex microphysical schemes have provided valuable insight to these processes, yet these models, and thus model forecasts, remain very sensitive to the microphysical settings used in the simulations (e.g., Fovell et al. 2009; Pattnaik et al. 2011; Khain et al. 2016; Brown et al. 2016). Accurate observations of microphysics in TCs are critical for constraining model simulations and understanding the involved processes. Unfortunately, such observations have traditionally been sparse and difficult to obtain.

Our current understanding of TC microphysics has been heavily shaped by relatively few in situ observations collected by precipitation probes on board aircraft penetrating a storm’s inner core. Several studies have focused on the distribution of ice particles. Black and Hallett (1986) found that TC convection contains high concentrations of ice above the 0°C isotherm, which leads to supercooled drops being quickly captured by ice and frozen upon contact. As a result, supercooled drops only occur above the −5°C isotherm in the strongest TC updrafts (Black and Hallett 1999). Houze et al. (1992) showed that the large ice particles in the inner core tended to be graupel within the eyewall near the formative updrafts and tended to be aggregates just outside the eyewall. Given these complexities of TCs, estimates of ice particle size distributions and ice water content (IWC) vary substantially around the storm (Black 1990; McFarquhar and Black 2004). Airborne and ground-based observations of raindrop sizes in TCs have also been critical for understanding TC microphysics, particularly by constraining the reflectivity–rain rate ($Z$–$R$) relationship for improved estimates of precipitation.
amounts (e.g., Merceret 1974; Jorgensen and Willis 1982; Ulbrich and Lee 2002; McFarquhar and Black 2004; Tokay et al. 2008).

One significant advancement in bulk microphysics observations has been the recent upgrade of the land-based WSR-88Ds in the United States to dual-polarization capabilities. The orthogonally polarized radar beams produce polarimetric variables that yield meaningful information on the types, shapes, and sizes of hydrometeors within the sampling volume (e.g., Doviak and Zrnić 1993; Bringi and Chandrasekar 2001; Ryzhkov et al. 2005a; Kumjian and Ryzhkov 2013a–c). Such dual-polarization radar observations have been extremely useful in advancing our microphysical understanding of various weather phenomena, particularly deep convective continental storms (e.g., Bringi et al. 1986; Zrnić et al. 1993; Hubbert et al. 1998; Ryzhkov et al. 2005b; Kumjian and Ryzhkov 2008; Homeyer and Kumjian 2015; and many others) and winter storms (e.g., Ryzhkov et al. 1998; Kennedy and Rutledge 2011; Andrić et al. 2013; Kumjian et al. 2013; Griffin et al. 2014b; Schrom et al. 2015; Moisseev et al. 2015; Schrom and Kumjian 2016; and many others).

In contrast, dual-polarization radar observations of TCs are comparatively sparse, yet observations that do exist have provided a great contribution to understanding TC microphysics. Using C-band dual-polarization radar, May et al. (2008) utilized a fuzzy logic scheme to classify the hydrometeor types within the inner core of Tropical Cyclone Ingrid. They found Ingrid’s precipitation structure to be highly asymmetric and identified midlevel graupel and hail in the convective regions of the asymmetric eyewall. Wang et al. (2016) examined microphysical characteristics of rainband precipitation in Typhoon Matmo after it made landfall. Using S-band dual-polarization radar, they retrieved drop size distributions that compared favorably with coincident disdrometer measurements. Griffin et al. (2014a) investigated the electrical and polarimetric radar characteristics of Tropical Storm Erin (2007) as it reintensified over Oklahoma. They found features similar to differential reflectivity (Z_{DPR}) columns that are common in continental convective storms [see Kumjian et al. (2014) and references therein] associated with the strongest updrafts around the eye, as well as depolarization streaks (Ryzhkov and Zrnić 2007; Kumjian 2013c; Hubbert et al. 2014) indicative of strong in-cloud electrification.

Although TCs are oceanic phenomena, landfalling or near-landfalling TCs captured by the upgraded polarimetric WSR-88D network would be an ideal step toward addressing the critical need for observations of TC microphysics. In an exploration of polarimetric rainfall estimation with upgraded WSR-88Ds, Ryzhkov et al. (2014) show data from Hurricane Irene (2011) in which retrievals of the normalized concentration parameter N_w (see Testud et al. 2001) revealed high concentrations of small drops, expected for tropical rainfall characterized by vigorous warm-rain processes. Brown et al. (2016) examined observations of Hurricanes Arthur (2014) and Ana (2014) using the upgraded WSR-88Ds. They compared these observations to simulated polarimetric radar variables using various microphysical parameterization schemes of the Advanced Research Weather Research and Forecasting (WRF) Model. Their results indicated a large discrepancy between the observed variables and the simulated variables, thus highlighting the critical need for improved microphysical representations in hurricane modeling. Recently, Kalina et al. (2017) also analyzed observations of Hurricanes Arthur and Irene and estimated the ice water path contributions from small and large ice particles in different regions of the storms. Also using the upgraded WSR-88Ds, Van Den Broeke (2013) detailed observations of biological scatterers within the eyes of Hurricanes Irene and Sandy (2012). These non-meteorological scatterers displayed clear signals in the polarimetric radar variables that could be detected and traced as each storm made landfall.

In the current study, we further examine WSR-88D dual-polarization observations of Hurricane Arthur. Our goal is to connect the microphysical processes gleaned from the polarimetric variables to the overall storm dynamics, as making this connection is key to advancing understanding of TC evolution and providing observational constraints for numerical simulations. We analyze observations in the context of their radial and azimuthal location within the storm. With this specific spatial analysis, we focus on the microphysics of the convective and stratiform precipitation for the various features within the storm.

Section 2 describes the radar observations and analysis methods. Section 3 examines the overall distributions of polarimetric variables throughout different regions of the storm. Sections 4 and 5 focus on observations above the melting level and section 6 presents the conclusions of this study.

2. Data and methodology

The polarimetric upgrade of U.S. National Weather Service WSR-88Ds was completed in June 2013. These S-band radars operate with scanning volumes comprising a series of surveillance scans at fixed elevation angles (0.5°–19.5°) that last 5–12 min, depending on the operational mode. The radar data utilized for this study
were observations of Hurricane Arthur collected during the time span of 0000 UTC 3 July–0600 UTC 4 July 2014 from three WSR-88Ds: KCLX (Charleston, South Carolina), KLTX (Oklahoma City, Oklahoma), and KMHX (Moorehead City, North Carolina). Figure 1 shows the track of Arthur along with the locations of each radar. Arthur formed from a tropical low off of the coast of central Florida that soon strengthened into a depression while remaining nearly stationary under weak steering flow. The depression began to rapidly intensify as it traveled to the north and then north-northeast (NNE). The storm became a hurricane as it entered the range of the KCLX radar at Charleston, South Carolina, radar. Arthur then intensified to maximum winds of 43 m s$^{-1}$ (category 2 on the Saffir–Simpson hurricane wind scale) and soon made landfall just west of Cape Lookout, North Carolina (Berg 2015).

Level-2 data from the WSR-88Ds were collected from all elevation angles for the analysis. The four variables examined in this study are the following: 

$Z_H$—reflectivity factor at horizontal polarization,

$Z_{DR}$—differential reflectivity,

$\rho_{HV}$—copolar correlation coefficient, and

$\Phi_{DP}$—propagation differential phase shift. Several measures were taken to ensure the quality of the analysis dataset. Slight biases in $Z_{DR}$ measurements from WSR-88Ds can exist. Thus, $Z_{DR}$ measurements from different radars were calibrated by comparing observations of the same precipitation features during this time. The resulting adjustment was a 0.05-dB increase to $Z_{DR}$ measurements from the KLTX radar. Relative trends in $Z_{DR}$ data, rather than absolute calibration, are of interest herein, and thus any small residual biases are inconsequential for our analysis. The data were then thresholded by $\rho_{HV}$ to remove nonmeteorological echoes: only signals with corresponding $\rho_{HV}$ values between 0.93 and 1.05 were retained for the analysis.¹

This study uses data collected out to 250-km range. As with all radars, the sampling volume increases as the beam spreads with increasing range. The WSR-88Ds have an angular beamwidth of 0.95°, which yields a ~1.6-km beamwidth at 100-km range and a ~4-km beamwidth at 250-km range. Given the scanning geometry of the WSR-88Ds, this beam spreading leads to degrading vertical resolution at large range. To account for the changing vertical resolution, the data were analyzed in two range groups: short range, which covers 0–100 km, and long range, which covers 100–250 km.

The data were next parsed into different regions relative to the storm center following the track of the storm throughout the dataset times. Best track data from the National Hurricane Center interpolated to hourly positions were used as an initial guess for the storm centers. These hourly centers were then nudged to a subjectively determined center position based on inspection of $Z_H$ and Doppler velocity fields from the closest observing radar. The storm centers were then interpolated to the starting time of each radar volume scan. We estimate that storm centers have an error of no more than 5 km; moreover, our results are not sensitive to variations of the storm center within this error range.

Hence and Houze (2011, 2012) analyzed the vertical distributions of precipitation in mature tropical cyclones by dividing the storm into annuli that separate the different precipitation features. We take a similar approach in our analysis. Eyewall convection at smaller radii is the driving engine of the storm, where large upward transport of high-$\theta_v$ boundary layer air leads to condensational heating that efficiently spins up the strongest winds of the storm (e.g., Emanuel 1986; Smith et al. 2008). Radially outward, convection is organized into spiral rainbands that shape the wind field at larger radii (e.g., Hence and Houze 2008; Didlake and Houze 2009). To best capture the precipitation features in Arthur, the selected annuli were 15–60 km for the eyewall (EW), 60–110 km for the inner rainbands (IR), and 110–200 km for the outer rainbands (OR).

¹ Note that $\rho_{HV}$ values >1 are unphysical and associated with erroneous corrections for low signal-to-noise ratio (e.g., Kumjian 2013a). Unfortunately, it is not possible to correct for these biased values given what is available in our level-II dataset. Including these echoes in our dataset did not impact our interpretations or conclusions.
The precipitation features are influenced by vertical wind shear of the surrounding environment. Through complex dynamical interactions, the environmental wind shear tends to align the eyewall and rainbands such that their convective and stratiform precipitation processes occur in predictable azimuthal regions of the storm (e.g., Chen et al. 2006; Hence and Houze 2011, 2012; Riemer 2016). Thus, the data were then separated into quadrants defined by the 850–200-hPa environmental wind shear vector. The deep-layer wind shear data come from the Statistical Hurricane Intensity Prediction Scheme (SHIPS) database (DeMaria et al. 2005). From this dataset, shear vectors were interpolated to each radar scan time, defining the four storm quadrants: downshear right (DR), downshear left (DL), upshear right (UR), and upshear left (UL). Figure 2 shows $Z_H$ fields from low-elevation radar scans and the 12 regions defined by the radial and quadrant divisions. Arthur clearly has an asymmetric distribution of precipitation in all annuli. Given these precipitation asymmetries, the WSR-88D scanning strategy, and the different region sizes, each region has a different number of data points that span different altitudinal distributions. We therefore analyze the bulk statistics of each region while taking into account the variances of data availability such that the regions can be fairly compared to each other. Table 1 displays the number of data points in each quadrant, annulus, and altitudinal layer analyzed in this study at both short range and long range.

3. Average of polarimetric variables

a. Profiles of $Z_H$

We begin by examining the $Z_H$ patterns that Arthur exhibited in each region. Figure 3 presents the average $Z_H$ vertical profile for all rings and quadrants. At short range, a local maximum in $Z_H$ associated with the melting layer bright band occurs near 4.5–5-km altitude, indicating the presence of stratiform precipitation in most rings and quadrants. At long range, the brightband signature is obfuscated given the coarser vertical resolution. Even with the bright band seen in the $Z_H$ average, convective precipitation may also be prominent in some annuli and quadrants and yield increased $Z_H$ values, particularly below the bright band. Variations in $Z_H$ are often found in TC eyewalls, indicating asymmetries in precipitation patterns that previous studies
have linked to the storm’s evolution. In the EW at both short and long range, the DR quadrant has the highest average $Z_H$ below 4 km, spanning 35–38 dBZ. Oppositely, the UL quadrant has the lowest average $Z_H$ below 4 km, spanning 28–33 dBZ. The DL and UR average $Z_H$ values lie in between for both short and long range. The average $Z_H$ suggests that convective precipitation is prevalent in the DR quadrant and stratiform precipitation dominates the UL quadrant.

Past studies have attributed an azimuthal variance in eyewall $Z_H$ to the interaction of the vortex core with the environmental wind shear. In this interaction, a wavenumber-1 eyewall asymmetry is produced where upward motion occurs in the downshear quadrants and downward motion occurs in the upshear quadrants. This asymmetry has been attributed to two dynamical processes: 1) a thermal balance response to shear-induced large-scale temperature gradients and shear-induced mesoscale tilt of the eyewall vortex (Jones 1995; Frank and Ritchie 1999; Reasor and Eastin 2012), and 2) a dynamic balanced response from differential vorticity advection between the lower and upper levels (Bender 1997; Frank and Ritchie 1999, 2001; Wu et al. 2006). Simultaneously, mesovortices embedded in the eyewall interact with the large-scale flow and concentrate low-level convergence and upward motion in the DR quadrant (Braun et al. 2006; Zhang et al. 2013; DeHart et al. 2014). The abundant and large hydrometeors generated in this quadrant are then advected downwind to the DL quadrant, where studies indicate that both a maximum of eyewall lightning and low-level $Z_H$ tend to occur (Marks and Houze 1987; Black et al. 2002;TABLE 1. Sample sizes of WSR-88D measurements (number of sampling volumes) for each annulus/quadrant region at both short and long range.

<table>
<thead>
<tr>
<th>Annulus Quadrant</th>
<th>Short range (0–100 km)</th>
<th>Long range (100–250 km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eyewall DR</td>
<td>$1.804 \times 10^6$</td>
<td>$2.740 \times 10^6$</td>
</tr>
<tr>
<td></td>
<td>$9.838 \times 10^6$</td>
<td>$6.365 \times 10^6$</td>
</tr>
<tr>
<td></td>
<td>$2.874 \times 10^6$</td>
<td>$3.228 \times 10^6$</td>
</tr>
<tr>
<td>Inner rainbands</td>
<td>$7.990 \times 10^6$</td>
<td>$6.010 \times 10^6$</td>
</tr>
<tr>
<td></td>
<td>$2.010 \times 10^7$</td>
<td>$6.852 \times 10^6$</td>
</tr>
<tr>
<td></td>
<td>$2.324 \times 10^7$</td>
<td>$3.941 \times 10^6$</td>
</tr>
<tr>
<td>Outer rainbands</td>
<td>$1.480 \times 10^4$</td>
<td>$7.764 \times 10^6$</td>
</tr>
<tr>
<td></td>
<td>$5.095 \times 10^7$</td>
<td>$2.031 \times 10^7$</td>
</tr>
<tr>
<td></td>
<td>$2.282 \times 10^7$</td>
<td>$1.543 \times 10^7$</td>
</tr>
<tr>
<td></td>
<td>$1.637 \times 10^6$</td>
<td>$5.897 \times 10^6$</td>
</tr>
</tbody>
</table>

Fig. 3. Vertical profile of average $Z_H$ in each storm region defined by annulus (eyewall, inner rainbands, outer rainbands) and shear-relative quadrant. Observations at (a)–(c) short range (0–100 km) and (d)–(f) long range (100–250 km).
The current observations have the strongest $Z_H$ echoes in the DR quadrant, which is displaced one quadrant upwind from these previous studies. However, a careful analysis shows that the maximum EW $Z_H$ shifts quadrants throughout the dataset time span, and the observations during the final 10 h after this shift contribute more to the analysis in Fig. 3, resulting in the DR maximum. Figure 4 displays the evolution of the average $Z_H$ in the eyewall region. Throughout most of the first 20 h (Figs. 4a–d), the maximum $Z_H$ for the lowest available data occurs in the DL quadrant, albeit the data availability limits our insight into the low-level $Z_H$ values. This DL maximum and overall azimuthal pattern (including a minimum of $Z_H$ in the UR quadrant) is consistent with previous studies, and thus shear is likely the dominant factor controlling Arthur’s eyewall asymmetry at these times.

In the final 10 h (Figs. 4e,f), the maximum $Z_H$ shifts to the DR quadrant. Several factors could contribute to such a shift. The first is a misalignment of the vortex tilt direction with the shear vector. The vortex tilt usually aligns just to the left of the shear vector (e.g., Reasor et al. 2004, 2013). But when the shear vector shifts suddenly, the tilt direction may take some time to adjust to the new shear direction, resulting in an eyewall asymmetry that aligns with the adjusting tilt direction rather than the shear (Stevenson et al. 2014). Analysis of NOAA P3 Tail Doppler radar observations indicate that, despite a changing shear direction, Arthur’s vortex tilt had shifted to the downshear direction by 2000 UTC (P. Reasor 2017, personal communication), which is very close to the expected alignment. Therefore, a misaligned vortex tilt does not appear to contribute to Arthur’s eyewall asymmetry.

Another possible factor could be the influence of storm motion. By 2000 UTC, Arthur’s movement had increased in magnitude (8 m s$^{-1}$) and turned to the NNE, while its environmental shear magnitude decreased to 3.5 m s$^{-1}$. Shapiro (1983) showed that a translating storm develops maximum boundary layer convergence in the front quadrants, which in turn produces enhanced precipitation just downwind. Frank and Ritchie (2001) and Chen et al. (2006) showed that when weak shear ($\sim 5$ m s$^{-1}$) is introduced, the eyewall asymmetry aligns more with the shear rather than the track motion. In the final 10 h of our case, the shear magnitude is less than that of Frank and Ritchie (2001) and Chen et al. (2006), and the track vector pointed...
30° to the left of the shear vector. This vector alignment would still predict a DL maximum in $Z_H$, whereas if the shear were too weak to be significant, previous studies predict a DL or UL maximum in $Z_H$. Neither case predicts a DR maximum, suggesting that the track motion does not solely contribute to the observed eyewall asymmetry.

Last, land interactions could influence the eyewall structure of a landfalling TC. Liu et al. (2007) suggested that in cases of weak environmental shear (<5 m s$^{-1}$), land effects on eyewall convection become more apparent. Some studies have found that a landfalling TC tends to develop an eyewall asymmetry with higher precipitation rates occurring near the onshore flow, which results from frictional convergence at the coastline (Tuleya and Kurihara 1978; Jones 1987; Gao et al. 2009). This expected land effect would predict a $Z_H$ maximum to occur in Arthur’s DL quadrant. On the other hand, other studies suggest that precipitation rates are largest near the offshore flow (Parrish et al. 1982; Blackwell 2000; Chan et al. 2004; May et al. 2008; Li et al. 2013). Chan and Liang (2003) found that these higher precipitation rates were a result of reduced static stability as dry air from the land was advected over moist air offshore. This effect would predict a maximum to occur in Arthur’s upshear quadrants. Neither onshore nor offshore enhancement fully explains the observed DR quadrant maximum.

We suspect that the eyewall asymmetry in the final 10 h comes from a complex interaction of wind shear, track, and land effects. By 2000 UTC, each of these plausible influences either emerged or notably changed magnitudes, and none of them individually explain the observed asymmetry. Still, it must be remembered that these observations are a single case study for which some variability from composite studies should not be surprising.

Unlike the EW region, neither the IR nor OR regions exhibited a notable azimuthal shift in the $Z_H$ pattern. Rather, their $Z_H$ profiles in Fig. 3 are representative of the reflectivity distribution throughout the dataset time span. The IR region exhibits the strongest $Z_H$ echoes below 4 km in the downshear quadrants and the weakest echoes occur in the upshear quadrants (Fig. 3) at both short and long range. These IR reflectivities are overall slightly weaker than those in the EW. The IR and OR $Z_H$ echoes require a careful examination because of a lack of short-range observations. The right-of-shear quadrants have a low data count compared to the left-of-shear quadrants (Table 1), which is a result of the radars’ scanning geometry, distance from the radar, and curvature of Earth. Consequently, some average $Z_H$ values may not be representative of the actual rainfall patterns. To account for this issue, we ignored locations where the data count at a specific altitude, quadrant, and range group makes up less than 1% of the total count in that annulus. Given the above sampling issue, the OR DR quadrant at short range is ignored and cannot be compared with other quadrants in this region. Despite this sampling issue, we find that the downshear quadrants in both IR and OR have higher reflectivities than the upshear quadrants at all altitudes. Past observational studies indicate that the DR quadrant in the rainband region contains the highest low-level $Z_H$ and most prominent convective precipitation (e.g., Didlake and Houze 2013a). The results here are in partial agreement. At short range, the DR quadrant in the IR has a slightly stronger $Z_H$ profile than the DL. At long range, the DR quadrant is stronger below the melting level and weaker than the DL above the melting level. In the OR at long range, the DL quadrant has a stronger $Z_H$ profile. With such mixed results, it is not clear which quadrant has the most convective precipitation from the $Z_H$ profiles alone.

For the remaining analyses of other variables ($Z_{DR}$, $\rho_{HV}$, and $\Phi_{DP}$), the time-composited plots were largely representative of the signal distributions throughout the last 20 h of the dataset (1000 UTC 3 July–0600 UTC 4 July). Other than the EW $Z_H$ analysis, there were no clear azimuthal shifts in the signal distributions for any of the annuli. In the first 10 h (0000–1000 UTC 3 July), the DR quadrant did not have enough samples in range to produce a representative signal at the pertinent altitudes (i.e., the quadrant data count was below 1% of the total data count; cf. Figs. 4a,b). Thus, the complete azimuthal evolution could not be adequately examined.

b. Profiles of $Z_{DR}$ and $\rho_{HV}$

Differential reflectivity $Z_{DR}$ is the ratio of the reflectivity factor at horizontal polarization to the reflectivity factor at vertical polarization, providing information on hydrometeor shapes. For particles small compared to the radar wavelength, larger $Z_{DR}$ values result from hydrometeors with a larger horizontal length relative to their vertical length, and $Z_{DR}$ values increase with a larger relative permittivity. The copolar correlation coefficient $\rho_{HV}$ measures the diversity of hydrometeor shapes, orientations, and relative permittivities within the sampling volume. For example, a distribution of homogeneous hydrometeors irrespective of size gives a $\rho_{HV}$ of unity. More shape diversity leads to decreased $\rho_{HV}$ values.

Figure 5 shows the average $Z_{DR}$ profile for each region at both short and long range. An overall trend in the $Z_{DR}$ profiles can be seen traveling from smaller to larger annuli. The quadrant profiles are most similar in the EW.
region, whereas the profiles are more spread apart at most altitudes in the IR region. In the OR, the spread increases further, particularly above 6-km altitude. Figure 6 presents the average $\rho_{HV}$ profiles at both short and long range. The same overall increase in spread is present in the $\rho_{HV}$ profiles, particularly between 3- and 8-km altitudes.

The observed similarities in quadrant profiles occur in the EW for two reasons: the EW region is smaller and the azimuthal winds here are stronger. Hydrometeors produced in a specific EW quadrant are more easily advected into the adjacent quadrant, given the stronger flow and shorter traveling distance between quadrants. In particular, the slowest falling ice crystals can be advected around the entire eyewall (e.g., Marks and Houze 1987), while ice crystals of intermediate fall speeds populate any EW quadrant along the way. With such azimuthal advection of hydrometeors, the EW quadrants exhibit hydrometeor populations that are more azimuthally homogeneous than that seen in the rainband regions, which have longer distances between quadrants and weaker advecting winds. We refer to this process, which produces homogeneity across quadrants, as azimuthal mixing of hydrometeor populations. The IR region has intermediate winds and traveling distances, and thus an intermediate spread between quadrants. Finally, the OR has the least azimuthal mixing of hydrometeor populations and thus has the most general spread between quadrants for the average $Z_H$, $Z_{DR}$, and $\rho_{HV}$ profiles.

The most prominent feature found in the average profiles (Figs. 5 and 6) for all annuli and quadrants is a pronounced $Z_{DR}$ maximum and $\rho_{HV}$ minimum associated with the melting layer. The $Z_{DR}$ enhancement results from slowly falling ice particles changing shape and relative permittivity as they melt in a shallow layer (e.g., Zrnić et al. 1993; Giangrande et al. 2008; Kumjian 2013b). The $\rho_{HV}$ minimum occurs where melting snow aggregates produce a large shape variety in a shallow layer due to their different melting rates, orientation angles, and liquid water fractions. In addition, aggregation within the melting layer could lead to variations in backscatter differential phase that also contributes to reduced $\rho_{HV}$ (e.g., Zrnić et al. 1993; Trömel et al. 2013, 2014). As with the $Z_H$ profiles, the $Z_{DR}$ and $\rho_{HV}$ enhancements associated with the melting layer are sharper peaks in the short range given the higher vertical resolution, while more rounded enhancements occur in the long range (with corresponding lower vertical resolution). A nearby sounding valid at 0000 UTC 4 July from KMHX indicates the 0°C level was at 4.8 km. This altitude matches the observed brightband enhancements.

![Fig. 5. As in Fig. 3, but for average Z_{DR}.](http://journals.ametsoc.org/doi/pdf/10.1175/MWR-D-17-0035.1)
in the inner rainband region. The melting level would increase some in the eyewall region. The brightband peaks in Fig. 5 show a ~0.5-km increase going from the outer rainbands to the eyewall.

Beneath the melting layer, $Z_{DR}$ and $\rho_{HV}$ are consistent with the differences in $Z_H$ found in each quadrant discussed above. At short range, the EW (Fig. 3a) has the highest $Z_H$ in the DR and UR quadrants, and the OR (Fig. 3c) has the highest in the DL and UL quadrants. The remaining panels in Fig. 3 have the highest $Z_H$ in the DR and DL quadrants. In the two quadrants where higher $Z_H$ values suggest heavier rain, we find larger $Z_{DR}$ and smaller $\rho_{HV}$ values, with the exception of $Z_{DR}$ in the long-range IR (Fig. 5e). In general, $Z_H$ and $Z_{DR}$ both increase with increasing rainfall rate (e.g., Cao et al. 2008; Kumjian 2013a), indicating larger, more oblate raindrops with increasing rain intensity. Heavier rain also tends to have a wider drop size spectrum, which slightly reduces $\rho_{HV}$. In contrast, lighter rain (indicated by lower $Z_H$ and $Z_{DR}$ values in Figs. 3 and 5) is associated with higher $\rho_{HV}$ (Fig. 6), because smaller drops are more spherical and drop size spectra tend to be narrower. Given that raindrop shapes do not vary widely (e.g., Pruppacher and Pitter 1971; Brandes et al. 2004; Thurai and Bringi 2005), $\rho_{HV}$ values do not drop below 0.98 at S band (Ryzhkov et al. 2005a). Also notable in the average profiles of $Z_H$ and $Z_{DR}$ at short range (Figs. 3 and 5, top row) is the increase in values toward the ground beneath the melting layer. This is the microphysical fingerprint of collision–coalescence growth of raindrops (Kumjian and Prat 2014), expected given the large warm (>0°C) cloud depth typical of TCs.

The DL and DR quadrants most consistently exhibit the heaviest rainfall throughout the storm, as seen in their $Z_H$, $Z_{DR}$, and $\rho_{HV}$ signatures. These downshear quadrants are also the expected locations of the heaviest rainfall in a sheared TC for both the eyewall (e.g., Black et al. 2002) and the rainbands (e.g., Hence and Houze 2012). These adjacent quadrants that span the azimuthal propagation of heavy rainfall also capture the evolution of TC convection at critical stages. In the subsequent analyses, we primarily focus on the DL and DR quadrants. To better understand their associated dynamics, we seek further evidence for convective and stratiform processes in the dual-polarization observations. This additional evidence can be found above the rain layer, where observations of the ice hydrometeors can provide further insight to the convection processes.

The brightband peaks of the DR quadrant have lower $Z_{DR}$ maximum values than those of the DL quadrant in all panels of Fig. 5. The DR $Z_{DR}$ peaks also occur either at the same altitude (Figs. 5a,f) or at a slightly higher
altitude (Figs. 5b,d,e) than those of the DL quadrant, which indicates a locally elevated melting layer.

Comparing the DL and DR $\rho_{HV}$ peaks, their relative values have a more mixed signal. In the short-range (Figs. 6a,b) and long-range EW (Fig. 6d), the DR $\rho_{HV}$ peak has a lower minimum value, while in the long-range rainbands (Figs. 6e,f), the DL $\rho_{HV}$ peak has a lower minimum value. Still, the relative shapes of the $\rho_{HV}$ profiles have a more consistent signal. The short-range (Figs. 6a,b) and long-range EW (Fig. 6d) profiles have a wider vertical distribution in the DR. The DL and DR peaks in the long-range rainbands (Figs. 6e,f) are more similarly shaped; still, a close look indicates that the DR peak is just slightly wider. Overall, the DR quadrant has a deeper layer of diverse hydrometeors between 3- and 7-km altitude in the short range and between 2 and 8 km in the long range.

The wider DR $\rho_{HV}$ peaks in the short-range panels (Figs. 6a,b) are attributed to a clear upward extension of lowered $\rho_{HV}$ values, jutting above the other quadrant profiles. At these altitudes (~4.5–6.5 km), $Z_{DR}$ values in the DR quadrant are also slightly larger than in the DL quadrant, resulting in a slightly larger $Z_{DR}$ minimum near 6.5-km altitude. Such differences in this layer could again indicate upward excursions of the melting layer. If the melting layer is actually at a steady altitude between 2 and 8 km in the long range.

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Farther aloft, there are additional notable differences in the $Z_{DR}$ and $\rho_{HV}$ average profiles. Although the DR and DL quadrants exhibit similar $Z_H$ values, the DR profiles feature larger $Z_{DR}$ and smaller $\rho_{HV}$ than DL. This is evident in the EW, short-range IR, and long-range IR and OR profiles. Larger $Z_{DR}$ values for a given $Z_H$ suggest a larger degree of crystal anisotropy in the DR than DL quadrants at this level, which corresponds to about ~15°C. This signature is similar to those of planar crystal growth regions observed in midlatitude precipitating systems (e.g., Kennedy and Rutledge 2011; Andrić et al. 2013; Bechini et al. 2013; Schrom et al. 2015; Moisseev et al. 2015; Schrom and Kumjian 2016; Kumjian and Lombardo 2017). At temperatures near ~15°C, plates are the primary crystal habit (e.g., Bailey and Hallett 2009). Greater supersaturations with respect to ice lead to branching and secondary habits like dendrites. Owing to their extreme aspect ratios, both plates and dendrites produce enhanced $Z_{DR}$ values. Thus, the difference in $Z_{DR}$ signatures between DR and DL suggests a difference in crystal growth in these environments. Specifically, the observations could mean that the DL has more subsaturated air, leading to sublimation and thus loss of crystal anisotropy, or that the DR has an environment more conducive to crystal growth because of higher moisture content. The latter is consistent with the presence of more active convection, as sufficient updrafts at these altitudes are necessary to maintain the needed supersaturation levels.

Previous observations suggest average updrafts of 2 m s$^{-1}$ can occur near 8 km in convective regions of the eyewall at these altitudes, whereas downdrafts at this level can be present in other eyewall regions (Reasor et al. 2013; DeHart et al. 2014). In the convective regions of rainbands, updrafts at the 8-km level can be 2 m s$^{-1}$ on average and 4 m s$^{-1}$ in an individual cell (Hence and Houze 2008; Didlake and Houze 2009), whereas stratiform regions have weaker (<1 m s$^{-1}$) updrafts on average (Didlake and Houze 2013b). These differences in vertical velocities around the storm can lead to the different moisture environments and the different crystal growth rates suggested by the observations.

We have discussed how the current observations from the eyewall have higher $Z_H$ values, usually an indicator
of convective precipitation, in the quadrant upwind of that shown in previous studies. These past studies also identify the entire downshear half of the eyewall as being more convectively active than the upshear half. But in Arthur, the observations indicate that the right-of-shear half of the eyewall is more convectively active. Both the DR and DL quadrants exhibited multiple signatures of active convection. Heavy rainfall is indicated by large low-level $Z_H$ values at later times of the observation period (Fig. 4f) and at short range (Fig. 3a). Planar crystal growth is apparent with similar observation period (Fig. 4f) and at short range (Fig. 3a). Planar crystal growth is apparent with similar

Compared to the UL quadrant with a similar flow) first overlaps with a sufficiently moist atmosphere. A vortex interaction with the low-level environmental region of preferred low-level convergence (created by cells are first generated in the UR quadrant where a vertical wind shear aligning both the moisture field and vortex tilt. In this alignment, rainfall convective cells are first generated in the UR quadrant where a region of preferred low-level convergence (created by vortex interaction with the low-level environmental flow) first overlaps with a sufficiently moist atmosphere. Compared to the UL quadrant with a similar $Z_H$ average, the UR quadrant exhibits several signatures that suggest an increased presence of convective processes: greater $\rho_{HV}$ values in the melting layer (Fig. 6) indicate more homogeneity of melting ice particles, lower $Z_{DR}$ maxima in the melting layer and smaller $Z_{DR}$ values in the layer just above (Fig. 5; seen in the short-range OR and long-range IR and OR) indicate a vertical stretching of the hydrometeor population distribution, and decreased $\rho_{HV}$ and increased $Z_{DR}$ values aloft (Figs. 5 and 6) indicate planar crystal growth.

Downwind in the DR quadrant, sufficient moisture and preferred convergence encourages further cell growth and new cell initiation, such that convectively active precipitation becomes more prevalent, which is again consistent with the increased $Z_H$ and the presence of the previously described $\rho_{HV}$ and $Z_{DR}$ signatures. Using lightning data from several storms, Corbosiero and Molinari (2002, 2003) identified the DR quadrant of rainband regions as the most robustly convective quadrant. Thus, the convective signatures identified in Arthur are consistent with their results.

Stratiform processes dominate in the left-of-shear quadrants, as all of the lighter, slowly falling ice particles generated by the convective region get adveected much farther downwind and concentrate in these quadrants. Using satellite radar data from several storms, Hence and Houze (2012) confirmed the stratiform nature of rainband precipitation in these quadrants. May and Holland (1999) and Didlake and Houze (2013b) examined the kinematics associated with the stratiform rainband precipitation also occurring in the left-of-shear half. Modeling studies also identify this azimuthal organization of the rainband complex (e.g., Moon and Nolan 2015). Based on these studies, the pronounced transition from convective to stratiform polarimetric signals between the DR and DL quadrants is consistent and supported by the expected vortex dynamics.

4. Analysis of layers above melting level in $Z_H$–$Z_{DR}$ space

Both $\rho_{HV}$ and $Z_{DR}$ distributions exhibited distinctive features within and above the melting layer that indicated a significant presence of convective microphysical processes in the DR quadrant when compared to the DL quadrant, which was likely dominated by stratiform processes. Smoothed and wider extrema within the melting layer extended upward to 6.5-km altitude in short-range observations and ~7.5-km altitude in long-range observations. The 6.5- and 7.5-km altitudes roughly correspond to the ~9° and ~14°C levels. Continuing upward, larger $Z_{DR}$ and smaller $\rho_{HV}$ values extended to 10-km altitude (~31°C) in the DR quadrant. Although signatures of each layer suggest the presence of active convection, these features point to different ongoing ice processes. We continue the examination of the different microphysical processes by focusing on these two layers above the melting level.

One useful method for examining the nature of precipitation using dual-polarization observations is by plotting on $Z_H$–$Z_{DR}$ space (e.g., Cao et al. 2008; Kumjian and Prat 2014; Brown et al. 2016). The reflectivity factor $Z_H$ is a function of both number and particle size, so a given value can correspond to different combinations of both. By examining joint $Z_H$–$Z_{DR}$ distributions, we can discriminate between differently shaped ice crystals at a given reflectivity, leading to an increased understanding of ongoing microphysical processes. We calculated the frequency of occurrence for all $Z_H$–$Z_{DR}$ combinations within each quadrant, ring, range group, and layer. Figure 7 shows the 50th percentile contours of the corresponding density plots for the lower layer. For the short-range observations, the lower layer
was selected as 5.0–6.5 km to capture the upward extension of the $Z_{DR}$ and $\rho_{HV}$ extrema (Figs. 5 and 6). For the long-range observations, 6–7.5-km altitude was chosen for the lower layer as this layer matches the points where the DR and DL $Z_{DR}$ average profiles diverge (Fig. 5).

Given the results from the previous section, we focus on the transition of the distributions between the DR and DL quadrants. In the EW short range, the $Z_H$ of the DR and DL quadrants both reach a maximum of 26 dBZ or higher, but the larger reflectivities of the DR are associated with $Z_{DR}$ values of up to 0.60 dB, about 0.10 dB higher than the DL. In the EW long range, the DR and DL quadrants have more similar distributions. Compared to the EW region, the $Z_H$–$Z_{DR}$ distributions in the rainband regions have a greater spread between the quadrants. This larger spread is again consistent with the decreased azimuthal mixing of hydrometeor populations between rainband quadrants. In particular, the DR and DL distributions have a notable difference. For the high-$Z_H$ and high-$Z_{DR}$ echoes in the short-range IR, $Z_{DR}$ decreases by $\sim0.15$ dB from the DR to DL quadrants. The long-range IR and OR distributions have a similar shift in $Z_{DR}$ along nearly constant $Z_H$ values. With reduced mixing compared to the EW, these distributions suggest that the DR and DL rainband quadrants are dominated by distinct processes in this lower layer.

At these altitudes, the expected ice hydrometeor types are mostly aggregates and/or graupel. These hydrometeors tend to produce near-zero $Z_{DR}$ across a range of $Z_H$ values. However, for sufficient ice supersaturations, nucleation and vapor depositional growth can occur. Columnar ice crystals have been observed in tropical cyclones (e.g., Black and Hallett 1986). Columnar habits are expected for vapor depositional growth at temperatures between $-23^\circ$ and $-28^\circ$C (e.g., Bailey and Hallett 2009; Lamb and Verlinde 2011), which roughly corresponds to 5.5–6.5-km altitude in the eyewall of Arthur. Given the relatively high temperatures in this layer, secondary nucleation (i.e., rime splintering; Hallett and Mossop 1974) is a more likely mechanism than primary nucleation for producing pristine columnar crystals. This requires supersaturation with respect to both liquid and ice and thus appreciable vertical motion to supply the excess vapor (e.g., Zawadzki et al. 2000).

The addition of columnar crystals in the presence of aggregates and graupel is not expected to increase $Z_H$ much, but has been shown to increase $Z_{DR}$ by several tenths of a decibel in scattering calculations (e.g., Vogel

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**Fig. 7.** Joint frequency distributions of $Z_H$ and $Z_{DR}$ for each storm region and defined altitude layers normalized by the maximum frequency in each region. Only the 50% contour lines are shown. The frequency bins are 1 dBZ for $Z_H$ and 0.2 dB for $Z_{DR}$. Observations (a)–(c) at short range taken between 5.5- and 6.5-km altitude and (d)–(f) at long range taken between 6.5- and 7.5-km altitude.
et al. 2015) and observations (Giangrande et al. 2016). Columnar ice crystals produced via secondary nucleation is a plausible explanation for increased $Z_{DR}$ values observed in the DR quadrant since upward motions tend to be vigorous here in both the eyewall and rainband regions (e.g., Braun et al. 2006; Hence and Houze 2008; Zhang et al. 2013; DeHart et al. 2014; Didlake et al. 2017). Additionally, intense updrafts can advect the columns farther upward, thereby increasing the $Z_{DR}$ at even higher altitudes (e.g., Hogan et al. 2002). However, this layer where columns form in the eyewall tends to be relatively thin given that they are aggregated quickly by the numerous ice crystals formed and falling within the eyewall (Black and Hallett 1986, 1999). Downwind in the DL quadrant, updrafts are generally weaker, so columns are not contributing as much to $Z_{DR}$ over this broad layer, which is consistent with the observed decrease in $Z_{DR}$ for larger $Z_H$. The ice column populations in this lower layer likely decreases in the rainband stratiform DL quadrant, whereas the EW region does not exhibit as much difference in the DR-to-DL $Z_{DR}$ decrease due to increased azimuthal mixing of hydrometeor populations.

We next examine the upper layer (~7.5–10-km altitude) where $Z_{DR}$ values exhibit notable differences between quadrants (Fig. 5) and $\rho_{HV}$ shows a slight trend toward lower values (Fig. 6), all for both short range and long range. These signatures were attributed to planar depositional growth near the ~15°C level. Figure 8 shows the $Z_H-Z_{DR}$ density plots for this upper layer. There are larger $Z_{DR}$ values for similar $Z_H$ values in the DR quadrant, particularly in the short-range EW and in the IR at both ranges. Similar to the lower layer, the enhanced $Z_{DR}$ values indicate that ice particles in the DR quadrant tend to be more anisotropic than those in the DL quadrant at altitudes above 7.5 km. Maximum $Z_H$ are slightly larger in this quadrant, as well. Taken together, these are suggestive of enhanced planar crystal growth, which again is consistent with increased convective activity in this storm quadrant.

In addition to increased $Z_{DR}$, the presence of pristine nonspherical crystals increases the propagation differential phase shift $\Phi_{DP}$. This has been documented both for secondary production of columns/needles between ~3° and ~8°C (e.g., Kumjian et al. 2016; Sinclair et al. 2016; Kumjian and Lombardo 2017) and platelike or dendritic crystals near ~15°C (e.g., Kennedy and Rutledge 2011; Andrić et al. 2013; Bechini et al. 2013; Schrom et al. 2015; Kumjian and Lombardo 2017). The variable $\Phi_{DP}$ is the difference in phase accumulation that occurs when horizontally polarized (H) and vertically polarized (V) radar waves propagate through nonspherical precipitation. When encountering hydrometeors with their maximum dimensions in the

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**Fig. 8.** As in Fig. 7, but for observations between 7.5- and 10-km altitude.
horizontal, the H wave acquires a greater phase shift than the V wave. As a result, $F_{DP}$ increases when such particles are present. Because pristine ice crystals like columns and plates tend to fall with their long axis aligned roughly horizontally, they contribute to positive $F_{DP}$. Though aggregates and graupel tend to contribute negligibly to $F_{DP}$, high concentrations of pristine crystals are sufficient to produce an observable signal, even at S band where phase accumulations are smaller. In the next section, we further examine the hypothesis that pristine ice crystals are responsible for the increased $Z_{DR}$ using $F_{DP}$ measurements.

5. $F_{DP}$ in the layers above the melting level

The $F_{DP}$ observations from dual-polarization radars tend to be noisy, particularly in ice (e.g., Bringi and Chandrasekar 2001). To address this issue, we utilize a bulk approach with the following procedure. An analysis volume within the hurricane is first defined by the shear-relative quadrant, the annulus, and an altitudinal range. For every radar wave (at each radar, elevation, and azimuth) that passes through this volume, $F_{DP}$ observations at every range gate of this beam segment are collected. Each beam segment is then normalized to begin with a range of zero and a $F_{DP}$ of 0 at the point that the beam enters the analysis volume. This is done to highlight the accumulation of $F_{DP}$ over the propagation path through the analysis volume rather than the absolute $F_{DP}$ values, which are affected by propagation through media uprange, and different radar system differential phases. We then create a frequency density plot where the normalized $F_{DP}$ values at each normalized range are collected, counted, and displayed. Finally, the $F_{DP}$ distributions are normalized by the total data count at each normalized range.

Figure 9 displays density plots of $F_{DP}$ for the short-range EW region between 5.0- and 6.0-km altitude. Each quadrant shows a notable spread in the $F_{DP}$ distribution indicative of the signal noise. Yet, the distributions indicate an overall increase in $F_{DP}$ with range. Such an increase indicates ice particles with their maximum dimension aligned in the horizontal in the hydrometeor population of the analysis volume. We quantify this increase by estimating the specific differential phase shift, or $K_{DP}$, which is equal to half of the range derivative of
This is done by averaging the normalized $\Phi_{DP}$ at each range, and using these values to calculate the slope over a selected range. We performed a Student’s $t$ test and found that the averages calculated at each range are statistically significant at the 99% confidence level. We examined data within a normalized range of 10 km to avoid the domain edges where the data count drops off, and the $K_{DP}$ estimates are shown as insets in Fig. 9.

Just as in previous analyses of dual-polarization measurements, the transition between DR and DL quadrants highlights a notable shift. The DR quadrant has $\Phi_{DP}$ values that increase at a faster rate than the DL quadrant. The $K_{DP}$ in the DR quadrant is 0.165 km$^{-1}$. Traveling into the DL quadrant, the $K_{DP}$ decreases by 0.06 km$^{-1}$.

If $Z_H$ remains constant for a given ice population dominated by nonpristine particles, an increased $K_{DP}$ should lead to a decrease in $\rho_{HV}$ because an increased presence of pristine (nonspherical) particles would enhance the hydrometeor diversity. While the DR and DL quadrants do not have similar $Z_H$ averages in the 5–6-km layer, the DR and UR quadrants do. With nearly identical $Z_H$, the $\rho_{HV}$ decrease going from the UR to DR quadrants is consistent with the corresponding $K_{DP}$ increase of $\sim 0.07$ km$^{-1}$. Continuing into the DL quadrant, the $Z_H$ average decreases. If the $K_{DP}$ decline corresponded to an equal proportion of pristine particles in the ice population, then the $\rho_{HV}$ would remain constant; however, the $\rho_{HV}$ average has a notable increase (Fig. 6). Therefore, the larger $K_{DP}$ in the DR quadrant indicates increased pristine ice crystal concentrations, which supports the hypothesis gleaned from the previous $Z_H$ and $Z_{DR}$ analyses. Figure 10 shows the density plots $K_{DP}$ for the short-range inner rainbands. The estimated $K_{DP}$ values shown here and the previous analyses of $Z_H$, $Z_{DR}$, and $\rho_{HV}$ tell a similar story for the DR–DL transition as in the short-range EW, which is also consistent with the expected convective evolution in these rainband quadrants.

To provide a fuller picture of the larger $\Phi_{DP}$ field, we used this same density plot analysis to estimate the $K_{DP}$ of 1-km layers up to the 9.5-km level (each altitude point represents the midpoint of the 1-km layer). Figure 11 presents the vertical profiles of estimated $K_{DP}$ resulting from this analysis for each region. The long-range $K_{DP}$ values were estimated over a normalized range up to 15 km. Changes in this normalized range upper bound did not produce significant deviations in the estimated $K_{DP}$ profiles.

In the short range, the higher $K_{DP}$ estimates of the DR quadrants seen in Figs. 9 and 10 extend over a shallow layer (5–6 km) and are part of local $K_{DP}$ maxima. Local $K_{DP}$ maxima in the DR occur in both this layer (5–5.5 km) and also at 7.5–8.0-km altitudes in the

![Figure 10](http://journals.ametsoc.org/doi/pdf/10.1175/MWR-D-17-0035.1)
short-range EW and IR. Distinct maxima are smoothed out in the long-range profiles, though the IR shows larger $K_{DP}$ values in DR compared to other quadrants throughout most altitudes. The maxima are consistent with large concentrations of pristine ice crystals; specifically, columns at the lower altitude and planar crystals aloft.

Assuming only columns contribute to the lower-altitude $K_{DP}$ signal, we can estimate their ice water content (IWC) following Ryzhkov et al. (1998). Observed bulk $K_{DP}$ values of $0.1^\circ$–$0.2^\circ$ km$^{-1}$ correspond to IWC of $\sim 0.3$–$0.6$ g m$^{-3}$. These values are broadly consistent with reported in situ observations in TCs (e.g., Black and Hallett 1986). This is also consistent with inferred columnar crystal concentrations at the same temperature ranges from other S-band observations in winter extratropical cyclones (Kumjian and Lombardo 2017). C-band observations suggest a maximum $K_{DP}$ of $0.2^\circ$–$0.3^\circ$ km$^{-1}$ for cases of secondary ice production, which (given the wavelength dependence of $K_{DP}$ for small particles like columns) corresponds to similar IWC values (Sinclair et al. 2016; Kumjian et al. 2016). In the latter study, aircraft observations confirmed the presence of $0.3$–$0.4$ g m$^{-3}$ IWC of needlelike crystals in such regions of enhanced $K_{DP}$. Electromagnetic scattering calculations by Kumjian et al. (2013) and Vogel et al. (2015) confirm that such IWC of columnar ice crystals in the presence of isotropically scattering particles like graupel or aggregates can lead to enhancements in $Z_{DR}$ of a few tenths of a decibel, depending on assumptions about columnar crystal aspect ratios and densities. Larger $K_{DP}$ values aloft in the planar crystal growth zone suggest larger IWC; applying the same formula for $0.2^\circ$–$0.3^\circ$ km$^{-1}$ corresponds to $\sim 0.6$–$1.0$ g m$^{-3}$. We are unaware of in situ measurements of pristine planar crystal IWC in tropical cyclones to compare with these estimates. Thus, though in situ microphysical observations in Arthur are not available, the polarimetric radar analysis is largely consistent with limited previous measurements and strongly suggests the presence of pristine ice crystals in large concentrations in the DR quadrant.

6. Conclusions

The WSR-88D network captured dual-polarization observations of Hurricane Arthur (2014) for 36 h as it made landfall in the eastern United States. These observations provided an opportunity to characterize the insufficiently understood microphysics of tropical cyclone (TC) precipitation. Our analysis focuses on the
microphysics of the convective and stratiform precipitation processes and puts them in the context of the overall TC dynamics. The data were separated by azimuth into wind shear–relative storm quadrants and separated by radius into eyewall (EW), inner rainband (IR), and outer rainband (OR) rings. The reflectivity $Z_H$ echoes exhibited a wavenumber-1 structure below the melting level, where the strongest echoes generally occurred in the downshear quadrants and the weakest echoes occurred in the upshear quadrants. The higher-resolution echoes at short range had a clear brightband signature associated with the melting layer in all quadrants, indicating the presence of stratiform precipitation; yet the $Z_H$ average profiles alone did not clearly characterize the prevalence of convective precipitation in each region.

The analysis of polarimetric variables more clearly discerns the likely ongoing precipitation processes. Vertical profiles of differential reflectivity $Z_{DR}$ and copolar correlation coefficient $\rho_{HV}$ in all regions displayed a local maximum of $Z_{DR}$ and a local minimum of $\rho_{HV}$ within the melting layer, corresponding to the $Z_H$ bright band, which are largely due to the changing relative permittivity and associated shape changes of melting particles. Below the melting layer, $Z_{DR}$ and $\rho_{HV}$ confirm signatures of heavy rainfall in the DL and DR quadrants; however, these two quadrants exhibited different shapes and distributions of polarimetric variables within and above the melting layer, suggesting that different precipitation processes contribute to the heavy rainfall.

The DR quadrant had multiple signatures of an increased presence of convective precipitation. Within and just above the melting layer, the $Z_{DR}$ and $\rho_{HV}$ profiles were more vertically stretched, exhibiting wider $\rho_{HV}$ brightband peaks, lower $Z_{DR}$ brightband maxima, and higher $Z_{DR}$ minima just above. The smooth, stretched profiles are attributed to intense, convective vertical motions, which spread the various hydrometeors across a large altitudinal layer. Above the melting level, these signatures indicate a significant variety of hydrometeor shapes, which includes a notable presence of nonspherical (more pristine) ice particles. These pristine ice particles are likely columnar ice crystals, as columns are the first ice crystals generated by convective upward motions at these levels.

Farther aloft, enhanced $Z_{DR}$ and decreased $\rho_{HV}$ indicated a second layer of enhanced pristine ice particles. At these levels near $-15^\circ$C, planar ice crystals rapidly grow in environments of higher moisture content, producing the observed polarimetric radar signatures. The presence of more active convection creates such an environment for rapid planar crystal growth, and this process occurring in the upper levels is consistent with the polarimetric signatures below which indicate convective precipitation in the DR quadrant.

The DL quadrant had signatures more consistent with predominant stratiform precipitation. These signatures included sharper $Z_{DR}$ peaks and narrower $\rho_{HV}$ peak widths at the bright band. Above the melting layer, such signatures indicate a more homogeneously shaped hydrometeor population, with smaller contributions of pristine ice particles. Aloft, $Z_{DR}$ and $\rho_{HV}$ profiles also indicated a decreased presence of pristine ice particles in the planar crystal growth zone. Stratiform precipitation with accompanying weak vertical motions produces such polarimetric signatures.

The locations of convective and stratiform precipitation are largely consistent with the expected orientation of tropical cyclone precipitation processes. Previous studies show that the most prevalent convective processes occur within the downshear (right of shear) half of the storm and stratiform precipitation inhabits the upshear (left of shear) half for the eyewall (rainband) region. This orientation is largely attributed to tilting of the vortex and organization of the moisture field by the environmental vertical wind shear.

The polarimetric signals in the eyewall quadrants shared more similarities than those in the rainband quadrants. Given that the eyewall regions are smaller and contain stronger azimuthal winds, hydrometeors were more readily mixed between quadrants, thus leading to azimuthal smearing of the observed profiles.

An analysis of propagation differential phase shift $\Phi_{DP}$ and specific differential phase shift $K_{DP}$ indicated two layers of increased ice crystals in the DR quadrant that match the columnar and planar crystal growth zones, as seen in the observations at short range. In these regions, normalized $\Phi_{DP}$ values had the greatest increase with range, and thus the largest $K_{DP}$ ($0.16^\circ - 0.20^\circ$ km$^{-1}$ in columnar growth zone and $0.55^\circ$ km$^{-1}$ in planar growth zone). Immediately downwind, the DL quadrant marked a sharp transition in the hydrometeor population, indicated by notably decreased $K_{DP}$, which were accompanied by decreased $Z_{DR}$, and increased $\rho_{HV}$. These features indicating a notable decrease in pristine ice crystals suggest a clear transition from convective to stratiform precipitation. Columnar and planar crystals grow in the DR convective regime, and then are advected downwind into the DL quadrant. Here, these ice particles aggregate and fall slowly and uniformly into the melting layer as part of the stratiform precipitation regime.

Identifying the distributions of dual-polarization measurements in TCs is a key first step in exploiting the new WSR-88D capabilities for TC research. Given...
that this was a single case study, the ubiquity of the results herein need to be examined by investigating observations of more storms as they enter the radar network range. To understand how the observed microphysical characteristics affect the overall storm structure and evolution, future research must identify the exact microphysical processes (and their thermodynamic impact) occurring in these dual-polarization observations. Tropical cyclone evolution is highly sensitive to the magnitude, location, and shape of heat sources and sinks within the storm vortex. The convective and stratiform processes identified in the current study indicate differences in the heating structures around the storm; future work must examine further details of these differences in order to connect these microphysical observations to the dynamics. Numerical models are an essential tool to make this connection. In a similar manner to Brown et al. (2016), future research should compare these results to numerical simulations using various microphysics schemes as part of the ongoing effort to improve hydrometeor representations in TCs.

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