Polarimetric Radar and Surface-Based Precipitation-Type Observations of Ice Pellet to Freezing Rain Transitions

DANA M. TOBIN AND MATTHEW R. KUMJIAN

Department of Meteorology and Atmospheric Science, The Pennsylvania State University, University Park, Pennsylvania

(Manuscript received 26 April 2017, in final form 12 September 2017)

ABSTRACT

Recent studies document a polarimetric radar signature of refreezing. The signature is characterized by a low-level enhancement in differential reflectivity $Z_{DR}$ and a decrease in the copolar correlation coefficient $r_{hv}$ within a region of decreasing radar reflectivity factor at horizontal polarization $Z_H$ toward the ground, called the refreezing layer (RFL). The evolution of the signature is examined during three winter storms in which the surface precipitation-type transitions from ice pellets to freezing rain. A modified quasi-vertical profile (QVP) technique is developed, which creates inverse-distance-weighted profiles using all available polarimetric data within a specified range from the radar location. Using this new technique reveals that the RFL descends in time prior to the transition from ice pellets to freezing rain and intersects the ground at the approximate transition time. Transition times are estimated using both crowdsourced and automated precipitation-type reports within a specified domain around the radar. These radar-estimated transition times are compared to a recently developed precipitation-classification algorithm based on Rapid Refresh (RAP) model wet-bulb temperature $T_w$ profiles to explore potential benefits of analyzing QVPs during transition events. The descent of the RFL in the cases analyzed herein is related to low-level warm-air advection (WAA). A simple method for forecasting the transition time using QVPs is presented for cases of constant WAA. The repeatability of the refreezing signature’s descent in ice pellet to freezing rain transition events suggests the potential for its use in operational settings to create or modify short-term forecasts.

1. Introduction and background

Freezing precipitation is a significant hazard in winter storms that can pose a great threat to a widespread region. Freezing rain creates dangerous conditions for motorists and pedestrians and can lead to increased accidents and falls due to slick roadways and sidewalks. Airline travel is particularly impacted by freezing precipitation both at the surface and in flight: flights may be delayed or canceled because of icing conditions on runways or aircraft wings, whereas in-flight icing situations where supercooled liquid drops freeze on contact with airframes can be particularly dangerous. Accumulation of such structural ice increases the weight of the aircraft, lessens the lift, reduces thrust, and increases drag, which is why aircraft icing has remained on the National Transportation Safety Board’s Most Wanted List of safety improvements since 1997 (Weener 2011; NTSB 2017). Ice accumulation on utility lines and tree limbs can cause them to be weighed down and snap, resulting in power outages and damage to property. Changnon and Creech (2003) indicated an annual average of $437 million (year 2015 U.S. dollars) in property-related losses from ice storms during a 50-yr period from 1949 to 2000. Since 1980, two severe ice storms categorized as weather and climate disasters produced damages exceeding $1 billion: an ice storm in January 1998 resulted in extensive forestry losses and 16 deaths in the northeastern United States with an estimated $2 billion in damages, and an ice storm in February 1994 resulted in 70 deaths in the southeastern United States with an estimated $5 billion in damages [year 2015 dollars; NCDC (2017)].

There are two pathways by which freezing rain or freezing drizzle may occur (e.g., Rauber et al. 2000 and references therein). One such mechanism is especially common in cases of freezing drizzle, in which so-called warm-rain processes occur at temperatures entirely below 0°C (e.g., Huffman and Norman 1988; Rauber et al. 2000). The second pathway begins with frozen hydrometeors melting as they descend through a layer of air with wet-bulb temperatures $T_w > 0^\circ$C. A shallow layer...
with $T_w < 0^\circ$C near the surface supercools these melted particles and enables them to freeze upon contact with the ground or elevated surfaces (e.g., Brooks 1920; Zerr 1997; Rauber et al. 2000; Stewart et al. 2015). Ice pellets may form in a similar manner where partly or fully melted particles completely refreeze prior to contact with the ground within a thicker and/or colder region of $T_w < 0^\circ$C near the surface (e.g., Hanesiak and Stewart 1995).

Despite similarities in the formation of ice pellets and freezing rain, the impacts of ice pellets can be significantly less detrimental than those associated with freezing rain (e.g., Zerr 1997). As such, it is critically important to distinguish between these two types of precipitation. Ralph et al. (2005) articulated an issue important to distinguish between these two types of precipitation. Ralph et al. (2005) articulated an issue important to distinguish between these two types of precipitation. Ralph et al. (2005) articulated an issue important to distinguish between these two types of precipitation.

The need for improved precipitation detection is not isolated to NWP models, but also extends to hydrometeor classification algorithms (HCAs) based on polarimetric radar observations. The polarimetric upgrade to the operational National Weather Service (NWS) Weather Surveillance Radar-1988 Doppler (WSR-88D) radar network was completed in June 2013 and provides the following variables: the radar reflectivity factor at horizontal polarization $Z_H$, the differential reflectivity factor $Z_{DR}$, the specific differential phase $K_{DP}$ (one-half the range derivative of the differential propagation phase shift $\Phi_{DP}$), and the copolar correlation coefficient $\rho_{hv}$. A description of these variables can be found in Kumjian (2013), as well as in Andrić et al. (2013) for the specific context of winter precipitation. At present, the operational HCA for WSR-88D data (Park et al. 2009) has only 10 hydrometeor classes that do not include ice pellets or freezing rain and thus is incomplete for winter precipitation. Several studies have attempted to identify these winter precipitation types through various methods, such as conventional radar reflectivity factor and Doppler velocities (e.g., Prater and Borho 1992), or polarimetric radar variables either with thermodynamic profiles (e.g., Schuur et al. 2012) or without (e.g., Thompson et al. 2014). The addition of polarimetric capabilities to the WSR-88D radar network, in conjunction with observed or model-analyzed thermodynamic profiles, has the potential to aid in the efforts to improve the detection of winter precipitation types (Schuur et al. 2012).

Although considerable efforts have been made to detect winter precipitation types with polarimetric radar, few studies have used polarimetric radar to document winter storms with transitioning precipitation types. Martner et al. (1993) intercepted a storm in upstate New York on 15 February 1992 with a NOAA X-band, dual-polarization, Doppler radar. Data were collected during the storm as it transitioned from overcast conditions to snow, to ice pellets, to freezing rain, and finally to rain. The event was poorly forecasted with misleading model guidance for the time at which freezing rain would affect the region. The authors called for data assimilation from advanced observational systems and the use of polarimetric radar to improve the forecasts of freezing rain events.

Martner et al. (1993) used the circular depolarization ratio (CDR), which is directly related to $Z_{DR}$ (Seliga et al. 1984). The analysis of periods of ice pellets revealed a reduction in $Z_H$ beneath the melting layer bright band owing to the dielectric constant of liquid reverting to that of ice, with no significant change in CDR noted. However, using the current operational polarimetric radar variables of $Z_{DR}$, $K_{DP}$, and $\rho_{hv}$, Kumjian et al. (2013) identified a unique signature indicative of hydrometeor refreezing during ice pellet events. This signature is characterized by enhancements in both $Z_{DR}$ and $K_{DP}$ and a reduction in $\rho_{hv}$ within a low-level region of $Z_H$ decreasing toward the ground called the refreezing layer (RFL). The RFL is distinct from the melting layer bright band aloft, which is characterized by increasing $Z_H$, $Z_{DR}$, and $K_{DP}$ and a marked decrease in $\rho_{hv}$. Differences in the melting and refreezing signatures, and a misalignment of variable maxima and/or minima, indicate complex microphysical processes associated with hydrometeor refreezing. Further, the refreezing signature’s evolution during precipitation transition events is unknown, as Kumjian et al. (2013) and other studies documenting the signature (e.g., Kumjian and
Schenkman 2014; R16; Van Den Broeke et al. 2016) focused only on times at which ice pellets were occurring.

In this study, we examine polarimetric WSR-88D radar data, surface-based precipitation-type reports from mpING and automated surface observing system (ASOS) data, and Rapid Refresh (RAP; Benjamin et al. 2016) model-analyzed thermodynamic profiles to determine how the refreezing signature characteristics evolve during transitions from ice pellets (IP) to freezing rain (FZRA). Radar observations show the refreezing signature descending in time and ultimately disappearing beneath the lowest radar-observable levels, coincident with when precipitation reports transition from IP to FZRA. We explore the potential advantages of using polarimetric data to verify or adjust high-resolution (e.g., RAP) model precipitation-type outputs by comparing them to polarimetric and precipitation-type observations. For transition events exhibiting constant low-level warm-air advection (WAA), we develop a simple forecasting technique that exploits the polarimetric features observed in the cases herein to forecast the IP to FZRA transition time.

Methods are detailed in the following section, followed by analyses of three transition cases in section 3. The proposed forecasting technique is detailed in section 4 along with synoptic features observed during the winter precipitation transition events. A brief discussion and conclusions of these analyses are presented in section 5, including suggestions for future work.

2. Analysis methods

Analyses of the IP-to-FZRA transition cases are based on the evolution of polarimetric radar data, RAP model output thermodynamic profiles, and all available local precipitation-type reports. Given the widespread nature of these precipitation events and the spatial distribution of precipitation reports, we have chosen a 50-km range from each radar location (i.e., a 50-km-radius circle) as our primary analysis domain. This choice, though admittedly subjective, is a balance between increasing the amount of available precipitation-type data and limiting the number of reports that are not truly representative of the polarimetric radar data, as explained below.

a. Polarimetric radar and range-defined quasi-vertical profiles

All radar data are from operational S-band polarimetric WSR-88D radars. The analysis will focus on $Z_H$, $Z_D$, and $\rho_v$: $\Phi_D$ and $K_D$ are omitted because typical maximum S-band $K_D$ values associated with the refreezing signature are $<0.1$ km$^{-1}$ (Kumjian et al. 2013), making it difficult to detect in an operational setting. Further, $K_D$ does not seem to provide significant additional information necessary to identify refreezing in these cases.

Herein, polarimetric radar data are presented as time series of what we call “range defined” quasi-vertical profiles (QVPs). This adaptation of the QVP method (Ryzhkov et al. 2016, hereafter R16) leverages the benefits of the existing technique while allowing for more generalized applications by lifting the restrictions set by using higher elevation angles. This modification is better suited for the present study and is in line with the QVP’s original intent of being “modified in the course of its future exploration” (R16).

A single QVP comprises the azimuthal median or mean of a plan position indicator (PPI) scan (Kumjian et al. 2013; R16), wherein range is converted into height above radar or ground level assuming standard beam refraction (e.g., Doviak and Zrnić 1993). QVPs typically are constructed with higher (>10°) elevation angles to minimize the area over which the variables are averaged at greater ranges (i.e., altitudes). However, the first eight 250-m range gates of WSR-88D radar data are censored, which can partially or fully mask low-level (<1 km) signatures such as refreezing when using higher elevation angles. Although lower elevation angles reduce the impact of data censoring, the area over which the data are averaged rapidly increases with range. For example, at a 19.5° elevation angle, data at 1 and 5 km above radar level are averaged over radii of ~3 and 15 km, respectively, whereas at 0.5°, data at 1 and 5 km correspond to averages over radii of ~75 and 230 km, respectively. Such a trade-off between low-level data censoring and QVP averaging area exists for all elevation angles.

Attempts to better resolve low-level polarimetric data with QVPs were discussed in R16 and Van Den Broeke et al. (2016). Both employed lower QVP elevation angles (6.4° and 2.4°, respectively). The choice of elevation angle depended on the application for the same 2 March 2014 central Oklahoma refreezing event: R16 chose the highest elevation angle for which the refreezing signature was not censored, whereas Van Den Broeke et al. (2016) chose an even lower elevation angle to capture the signature over a larger area to match surface

---

1 Kumjian et al. (2013) used the median value, whereas R16 suggested “azimuthal averaging.” Quasi-vertical profiles computed herein are mean values.

2 Standard beam refraction was shown in Doviak and Zrnić (1993) to be a reasonable assumption in even severe temperature inversions, in which errors in beam altitude are less than a few hundred meters.

3 For example, at the highest elevation angle (19.5°), these censored range gates correspond to the lowest ~700 m in height above the radar. In contrast, only the lowest ~20 m are censored at the lowest operational elevation angle (0.5°).
precipitation-type reports. However, use of such low elevation angles tends to smear the features aloft, including the melting layer. R16 suggested using multiple, close elevation angles to construct a QVP to further reduce statistical uncertainty, but there would remain subjectivity with how many and which elevation angles to choose.

To mitigate these issues of elevation-angle choice, low-level censoring, vertical resolution, and increasing averaging area, we develop range-defined QVPs that utilize data from all available elevation angles. First, we define a range from the radar within which the data will be used. Then, traditional QVPs are computed for all available elevation angles. These QVPs are each interpolated onto a common vertical axis with 2-m resolution. For each height above the ground on the common axis, data from all elevation angles are averaged according to the specified range using an inverse distance weighting:

\[
W_i = \frac{1}{(r_i - (d - 1))^p},
\]

where \(v_i\) and \(r_i\) are the corresponding values and ranges of each QVP, \(d\) is the specified range (50 km is used herein), and \(p\) is a power parameter \(\geq 0\). Setting \(p = 0\) is equivalent to the simple mean of all values, whereas higher values result in a profile weighted most heavily to the value closest to the desired location. Here, we desire \(w_i = 1\) for all \(r_i \leq d\), and \(w_i < 1\) for all \(r_i > d\). To accomplish this, we set \(p = 0\) for all \(r_i \leq d - 1\), and \(p = 2\) for all \(r_i > d - 1\). For \(d - 1 < r_i \leq d\), \(w_i = 1\) such that equal weight is given to values within \(d\), and the weighting function drops off as the distance squared outside of \(d\); for example, \(w_i < 0.01\) after 59-km range. Thus, although a single QVP averages over an increasingly larger area with height, the area of range-defined QVPs is largely confined to the area set by the specified range. A conceptual depiction of the range-defined QVP geometry is shown in Fig. 1. Individual QVPs and the resulting range-defined QVP for data collected at 0001 UTC 4 January 2015 by the Albany, New York (KENX), WSR-88D radar are plotted in Fig. 2 to illustrate the method. Note that important features (e.g., the low-level refreezing signature and the overlying
melting-layer signature) are reasonably resolved by the range-defined QVP.

The versatility of range-defined QVPs is that users may define a region over which the profiles are constructed. A smaller or larger value for \( d \) may be appropriate for different weather events or locations, though we found little sensitivity of the resulting profiles to \( d \) ranging from 30 to 70 km (not shown). In principle, one could also define two azimuths and two ranges to isolate a particular location, with \( w_j \) between the two ranges and \( w_j < 1 \) outside. This may be especially useful in cases of spatially heterogeneous or widely scattered precipitation, which makes the QVP technique less reliable (e.g., R16; Kumjian and Lombardo 2017). Similarly, this type of approach can be applied to separate convective from stratiform precipitation, as suggested by R16. Because data from all available elevation angles are used, range-defined QVPs are adaptable for any VCP. This is beneficial for events during which the VCP is changed (e.g., Kumjian and Lombardo 2017) and means range-defined QVPs will be unaffected by the upcoming VCP changes to WSR-88D radars scheduled for deployment in 2017 (NWS 2015).

b. Precipitation-type data

Although the 50-km radius identifies the range for our range-defined QVPs, the precipitation data domain is extended to a range of 59 km to account for data included in the polarimetric profiles that are just beyond 50 km, yet still contribute to the profile with a weighting function \( w_j > 0.01 \). This 59-km radius precipitation analysis domain captures an adequate number of precipitation-type reports corresponding to the polarimetric data profiles. All available surface-based precipitation-type reports available within this domain are used. The inclusion of both mPING and ASOS data increases the number of reports available throughout each event and mitigates the potential limitations of each, as discussed below.

The primary precipitation-type data used for analysis are crowdsourced mPING reports. Because operational S-band radars are typically displaced from residential areas to provide coverage of these regions without being located within a residential neighborhood, it is unlikely for a large number of mPING reports to be submitted close to the radar. The 59-km radius was deemed appropriate after examining the distribution of reports for each case: a smaller range significantly reduces the number of reports included in the analysis, whereas a larger grid potentially includes reports not representative of the range-defined QVPs.

There are several limitations of utilizing crowdsourced mPING reports (see Elmore et al. 2014). The spatial and temporal resolution of reports is highly
dependent upon demographics, with more reports typically submitted in urban versus rural areas and during the day versus at night. Additionally, reports are not screened for accuracy. It is especially likely in mixed and/or transitional precipitation events that a precipitation type may not be correctly identified (Reeves 2016). For example, a hail report in the middle of a winter precipitation event is likely IP misidentified by citizen-scientists. Also likely to occur is the misidentification of IP as either SN or FZRA because of varying definitions of both “ice pellets” and “sleet” among users and industries (Reeves 2016). Despite these limitations, mPING reports are nonetheless found to be of adequate quality for research (e.g., Elmore et al. 2014, 2015). Further, for each of the cases herein, IP were observed personally by the authors, close confidants, or NWS employees, improving confidence in these reports.

Reeves (2016) assessed biases associated with both mPING and ASOS data during wintertime precipitation events, including a significant bias in mPING reports toward ice pellets and away from rain. This bias toward IP is suggested to be the result of mPING users being more likely to submit reports of IP if a mixture is occurring, and more likely to notice and report IP than the more commonplace RA during single-precipitation-type occurrences. In contrast, automated precipitation-type sensors are unable to identify IP or mixtures of precipitation. IP conditions often occur as mixtures with other forms of precipitation (e.g., Hanesiak and Stewart 1995; Cortinas et al. 2004) that the sensors can identify, which can result in SN or FZRA reports that do not indicate the additional presence of IP. However, trained observers at some instrument locations can correct errors and augment reports to include all precipitation types. The inclusion of both mPING and ASOS reports from all locations within the analysis domain helps to mitigate these biases and improve our confidence in the surface precipitation types. These reports are plotted separately beneath range-defined QVP plots, with precipitation types denoted by the abbreviations and/or symbols included in Fig. 3a. For reference, Figs. 3b–d contains the locations of all mPING reports and ASOS stations relative to the radar for each case.

c. Thermodynamic data

Thermodynamic profiles are obtained from hourly RAP analyses and forecasts. Because ice–liquid phase changes are contingent on \( T_w \), \( T_w \) is computed from model output temperature and relative humidity (RH) values following Stull (2011). The \( T_w \) contours are overlaid on all range-defined QVP time plots. We adopt the term “isopsychotherm,” hereby defined as a line of equal or constant \( T_w \).

3. Observational analysis

This section documents three IP-to-FZRA transitions using range-defined QVPs, surface precipitation reports, and RAP model analyses. These cases were selected during the course of a study of post–polarimetric upgrade winter storms that used mPING precipitation reports and WSR-88D radar data to identify IP storms characterized by pronounced polarimetric refreezing signatures. A handful of storms met those criteria, yet these select cases stood out as having a unique and repeated evolution of the RFL through the event. Although representing a limited number of cases, the analyses examined will demonstrate how one may assess confidence in precipitation-type transition forecasts within a 50-km-radius region around a WSR-88D radar location. Additional cases will be necessary to fully determine the utility of the range-defined signatures in an operational setting.

a. 3–4 January 2015

A winter storm transition event occurred on 3–4 January 2015, impacting Albany and the surrounding region. ASOS data for this event come from the station at Albany International Airport (ALB) northwest of KENX (Fig. 3b). Though light precipitation existed in the domain prior to 1900 UTC, the first precipitation reported by the automated system was just after 1900 UTC, coincident with mPING reports of SN over the next several hours. Between approximately 1930 and 2130 UTC, planar crystal growth is evident at 3.8–5 km near the \(-15^\circ C\) isopsychotherm with enhanced \( Z_{DR} > 1 \) dB and reduced \( \rho_{hv} \) (\(-0.9\) values within a region of low \( Z_{HI} \) \(<10\) dBZ; e.g., Kennedy and Rutledge (2011); Andrič et al. (2013); Bechini et al. (2013); Moisseev et al. (2015); Schrom et al. (2015); Kumjian and Lombardo (2017)]. Although not as evident as this earlier period, this signature persists throughout the event with \( Z_{DR} \) values \(>0.5\) dB and, again, peaks in intensity toward the end of the event (Fig. 4b).

---

4Hail has, however, been documented in winter precipitation events featuring ice pellets (e.g., Kumjian et al. 2013; Van Den Broeke et al. 2016).

5Following conventions in the AMS Glossary of Meteorology for lines of constant temperature (isotherm) and dewpoint temperature (isoschotherm), it is derived from the Greek terms \textit{isos} (meaning equal), \textit{psukhos} (meaning cold), and \textit{therme} (meaning heat).
Reports of IP/SN and IP were submitted to mPING as early as 2146 UTC, while ALB reported a mixture of IP and SN that began at 2229 UTC. This delay can be accounted for by the fact that ALB is ∼30 km northwest of KENX. The inclusion of IP reports is associated with the emergence of both a melting and refreezing layer in the range-defined QVPs after approximately 2200 UTC. These signatures are well defined at approximately 2 and 0.5 km, respectively, between 2300 and 0100 UTC (Figs. 4a–c), a period during which both mPING and ASOS data reported primarily IP, with a few additional reports of IP/SN and SN (Figs. 4d,e).

The first mPING report of FZRA was submitted at 0140 UTC, accompanying a report of IP and a report of RA/IP within the domain. Augmented ASOS observations at ALB indicated a mix of FZRA and IP at 0141 UTC. Thus, the mPING and ASOS observations are consistent in indicating that the IP-to-FZRA changeover began at ∼0140 UTC. The polarimetric refreezing signature in the range-defined QVPs of $Z_{DR}$ appears to have descended from its initial height (∼0.5 km ARL) prior to this transition. At the time of the transition (0140 UTC), the RFL has nearly disappeared, having descended to the lowest tens of meters above the surface, with only a slight $Z_{DR}$ enhancement visible in the lowest portions of the range-defined QVPs, potentially in association with the top of the RFL. This remaining portion of the RFL likely indicates that some hydrometeors were still freezing aloft while others remained unfrozen until contact, producing the mixture of

---

**Fig. 3.** Abbreviations and symbols (a) for both mPING and ASOS precipitation-type reports. The location of all mPING reports (symbolized) and ASOS stations (yellow squares) for (b) the 3–4 Jan 2015 event at Albany, (c) the 17 Feb 2015 event at Wakefield, and (d) the 3–4 Mar 2015 event at State College. Locations are centered on each respective WSR-88D radar location (black circle) with both the 50-km range-defined QVP and 59-km precipitation analysis domains (green and blue, respectively) denoted in (b)–(d).
IP and FZRA. Further, when only FZRA and RA were reported by ALB (after 0316 UTC) the $Z_{\text{DR}}$ enhancement had completely disappeared.

Figure 4 reveals an apparent association between the low-level $Z_{\text{DR}}$ maxima and the RAP-derived $-5^\circ\text{C}$ isopsychrotherm. The $Z_{\text{DR}}$ enhancement appears to be well defined at a height just below the $-5^\circ\text{C}$ isopsychrotherm between 2200 UTC 3 January and 0100 UTC 4 January. Interestingly, $-5^\circ\text{C}$ is also the $T_w$ threshold used for IP in the background classification algorithm for Schuur et al. (2012). However, Reeves et al. (2016) found that a $-6^\circ\text{C}$ $T_w$ threshold in a spectral bin microphysics model performs the best in reproducing the observations of IP.

A simple precipitation classification scheme based on the hourly RAP data is used to estimate precipitation type and to provide a model-output-derived approximation for the changeover time. This scheme follows the background classification in Schuur et al. (2012), who utilized output from the Rapid Update Cycle (RUC; replaced by RAP on 1 May 2012). Profiles with $T_w < 0^\circ\text{C}$ throughout the layer correspond to SN. Profiles with surface $T_w < 3^\circ\text{C}$ that have a single $T_w > 0^\circ\text{C}$ layer may be classified as IP if the maximum wet-bulb temperature in the profile $T_{w_{\text{max}}} < 2^\circ\text{C}$ and the profile minimum wet-bulb temperature $T_{w_{\text{min}}} < -5^\circ\text{C}$. If $T_{w_{\text{max}}} > 2^\circ\text{C}$ and $T_{w_{\text{min}}} \geq -5^\circ\text{C}$, precipitation is classified as FZRA. Hourly analysis profiles from 1900 to 0400 UTC are plotted in Fig. 5 with line colors corresponding to precipitation classifications using this scheme. This analysis shows a transition from SN to IP occurring between 2200 and 2300 UTC, and from IP to FZRA between 0100 and 0200 UTC. These times are both consistent with the estimates from surface reports and QVPs, indicating that the RAP likely handled this event well. Thus, polarimetric radar observations, surface precipitation-type reports, and RAP analyses all depict a coherent transition event. Consistency between these disparate data sources leading up to and throughout the transition event.
could boost forecaster confidence in short-term prognoses of precipitation type.

b. 17 February 2015

A winter storm affected several states from the southern plains to the mid-Atlantic on 16–17 February, producing significant accumulations of IP, SN, and FZRA. On 17 February 2015, the Wakefield, Virginia (KAKQ) WSR-88D radar displayed a pattern of evolution in the QVP refreezing signature similar to those shown in the previous case (Fig. 6). Unfortunately, there were several hours at the end of the event in which no mPING reports were submitted in the domain; however, ASOS and several local storm reports (LSRs) from the NWS at Wakefield are available. The ASOS at Wakefield (AKQ) is located right next to the KAKQ radar, and the station at Newport News/Williamsburg International Airport (PHF) is to the east (Fig. 3c).

Precipitation began as snow around 1800 UTC 16 February within the domain (not shown). Again, we see evidence of planar crystal growth above 3.8 km near the −15°C isopsychrotherm between 0400 and 0430 UTC, as well as enhanced $Z_{DR}$ values (>0.5 dB) above the melting layer (ML) that persists throughout much of the event. Prior to the melting layer appearing at 0330 UTC 17 February, low $Z_{H}$ values (<25 dBZ) with near-zero $Z_{DR}$ and high $\rho_{hv}$ indicate that dry snow aggregates likely were falling (Figs. 6a–c). At 0416 UTC, hints of an RFL emerge at 0.8 km with a slight enhancement in $Z_{DR}$. The $Z_{DR}$ enhancement strengthens in time, and the signature becomes clear by 0445 UTC. The Wakefield NWS office indicated that a transition to IP occurred at 0425 UTC with “moderate sleet at observation,” noted in an LSR. AKQ indicates a transition from light snow to precipitation of unknown type (UP) at 0428 UTC in the remarks section of the METAR, likely a reflection of this transition to IP. Farther east, PHF notes this transition from SN to UP occurring at 0500 UTC.

Between 0500 and 0700 UTC, the refreezing signature descended steadily from approximately 0.8 down to 0.4 km AGL. After 0700 UTC, however, the RFL remained between 0.2 and 0.4 km AGL for several more hours. While no mPING reports were submitted after 0700 UTC in the domain, local ASOS continued to report UP. ASOS precipitation-type sensors discern RA and SN based partly on fall speeds and assign UP when particle fall speeds do not correspond well with those typically associated with RA or SN. Although IP typically have fall speeds on par with RA, Nagumo and Fujiyoshi (2015) have documented some IP fall speeds to be slower, between the ranges of RA and SN. Thus, both mPING and ASOS reports are suggestive of a prolonged IP event occurring within the domain.

At approximately 0800 UTC, precipitation began to taper off with echoes becoming more isolated as inferred from PPI scans (not shown). Although range-defined QVPs reveal a decrease in the precipitation intensity with decreasing $Z_{H}$ values, the profiles are not able to indicate whether precipitation is homogeneously spread or scattered within the radar domain. A PPI taken at 0910 UTC (Fig. 7) shows scattered precipitation over half of the radar site with a heavier banded region southeast of the radar near the North Carolina coast. QVPs are designed for widespread precipitation; thus, our range-defined QVPs after 0800 UTC become increasingly unreliable as the precipitation gradually becomes more scattered. As a result, it is difficult to utilize these profiles to identify if a transition in precipitation type occurred for this event during this time. Future improvements to the QVP technique could include a reliability measure (e.g., how many azimuths are contained in the average) to more quantitatively assess the QVP reliability.

AKQ reports indicate UP from 0441 to 1023 UTC, followed by a period of precipitation types alternating between FZRA and UP from 1023 to 1254 UTC (Fig. 6e). PHF similarly begins reporting FZRA at 1042 UTC, shortly after the first reports of FZRA by AKQ (Fig. 6f). With these limited data, we speculate that a transition from IP to FZRA occurred within the domain at approximately 1030 UTC. However, both ASOS
locations report little to no accumulations after approximately 1000 UTC, and precipitation stopped altogether at 1303 UTC at AKQ and at 1339 UTC at PHF. Although it is evident that a transition in precipitation has occurred in the domain, the tapering in intensity and isolated nature of the remaining precipitation make it unclear in both the polarimetric and mPING data, highlighting the necessity of widespread precipitation for a QVP-based analysis.

Here, we examine the RAP data to determine if the model has captured the IP-to-FZRA transition. The shallow layer of $T_w > 0^\circ$C between approximately 0520 and 0900 UTC (Fig. 8) causes the Schuur et al. (2012) algorithm to identify IP as the likely precipitation. The timing for the SN-to-IP transition is about 30 min later than the observations discussed above, and the algorithm applied to RAP data indicates a transition back to snow just after 0900 UTC. The RAP-analyzed $0^\circ$C isopsychrotherm is poorly aligned with the melting layer that persisted from 0330 UTC through the end of the event. Additionally, the RAP does not exhibit any $T_w > 0^\circ$C after 0900 UTC, despite the clear presence of the melting layer in the radar observations. Although less obvious in $Z_H$ QVPs because of the scattered nature of the precipitation, the melting layer is evidenced by the persistent $Z_{DR}$ enhancement and decreased $\rho_{HV}$ values at 2–3 km AGL. Griffin et al. (2014) and Kumjian and Lombardo (2017) show similar cases where a melting layer is present in the polarimetric radar data, but the RAP model data do not indicate $0^\circ$C temperatures over the radar.

In contrast with the previous case where there was a close relation between the height of the $Z_{DR}$ enhancement associated with refreezing and the placement of the $-5^\circ$C RAP-derived isopsychrotherm, the $-5^\circ$C isopsychrotherm here is displaced $\sim 0.3-0.8$ km above the $Z_{DR}$ enhancement. It remains unclear whether this difference is the result of model errors, or if refreezing has additional dependences on other environmental variables. For example, Kumjian and Schenkman (2014) and

---

**Fig. 6.** Range-defined QVPs of (a) $Z_H$, (b) $Z_{DR}$, and (c) $\rho_{HV}$ obtained from the S-band KAKQ radar from 0200 to 1200 UTC 17 Feb 2015, with accompanying precipitation reports from (d) mPING and the (e) AKQ and (f) PHF ASOS locations. RAP model analysis derived isopsychrotherms at $5^\circ$C intervals are plotted and labeled in (a)–(c).
Nagumo and Fujiyoshi (2015) provided observations that evaporative cooling in dry air can increase the likelihood of refreezing; however, more research involving detailed microphysical modeling is required to determine the environmental parameter space conducive for refreezing. This case demonstrates the challenges in identifying precipitation transitions with limited surface reports, model errors, and diminishing precipitation intensity. Nonetheless, the polarimetric radar data and limited surface reports do provide some insight into the changing low-level thermodynamic conditions and the timing of the transition and should alert forecasters to model errors.

c. 3–4 March 2015

On 3–4 March 2015, a transition event impacted the northern and central regions of Pennsylvania and was well documented by the polarimetric WSR-88D radar near State College, Pennsylvania (KCCX), and surface precipitation reports (Fig. 9), including some made by the authors (Fig. 10). Precipitation reports include the manned AWOS station in State College (UNV) east of the radar (Fig. 3d) and the ASOS station in Clearfield (FIG) to the northwest (Fig. 3d). Unfortunately, only the following RAP analysis hours are available during this event: 1300, 1800, 2100, 0000, and 0400 UTC. The missing data are filled in with the previous available analysis cycle’s forecast.

Precipitation reports from all sources indicate SN occurring from 1454 to 1705 UTC (Fig. 9; 1454–1600 UTC). Precipitation reports from all sources indicate SN occurring from 1454 to 1705 UTC (Fig. 9; 1454–1600 UTC).
As with the previous cases, planar crystal growth zones are visible in the range-defined QVPs above the ML (at ~3–5 km AGL) with enhanced $Z_{DR}$ values (>1 dB) and reduced $\rho_v$. The $Z_{DR}$ maxima associated with the planar crystal growth region in this case appears to be at a lower altitude from the RAP-derived $-15^\circ C$ isopsychrotherm, potentially indicating a warm bias in the RAP model analyses.

The first indication of a transition from SN to IP occurred just before 1700 UTC, with FIG reporting UP at 1655 UTC and an mPING report of IP/SN submitted at 1658 UTC. However, a variety of precipitation types were reported between approximately 1700 and 1900 UTC, with mPING reports of IP, SN, FZRA/freezing drizzle (FZDZ), RA/drizzle (DZ), and mixtures thereof occurring within the domain while FIG continued to report UP, and UNV received no precipitation. During this time, range-defined QVPs show a distinct melting layer forming, although the presence of an RFL is unclear. This assortment of precipitation types is likely the result of both a shallow and/or weak melting layer and reduced precipitation rates within the domain.

At 1900 UTC, increased low-level $Z_H$ and $Z_{DR}$ values correspond to an increase in precipitation intensity within the domain. At this time, a $Z_{DR}$ enhancement is located approximately 500 m above the surface, suggestive of refreezing. However, it is unclear if all hydrometeors had fully frozen, as there is only a shallow
region near the surface with $Z_H$ and $Z_{DR}$ values decreasing toward the ground, indicative of the bottom of the RFL (Kumjian et al. 2013). Between 1900 and 0000 UTC, mPING reports were largely either RA/IP or IP, whereas FIG reported both FZRA and UP (note this location does not have a human observer). A trained observer at the UNV AWOS reported hourly precipitation types during this time. It is unclear when precipitation began at UNV as there are no indications in the METAR remarks, but observations show a prolonged period of IP for several hours prior to 0000 UTC 4 March and a single report of FZRA at 2053 UTC, all within the time during which mPING reports are either IP or IP/RA.

We rely on precipitation reports from mPING and UNV to estimate the IP-to-FZRA transition time, as FIG began reporting FZRA earlier in the event when other sources indicated IP as a primary precipitation type. FZRA was first indicated at 2353 UTC by an mPING report and at 0033 UTC 4 March at UNV. This 40-min discrepancy makes it difficult to identify an exact transition time; however, there is another mPING report of FZRA at 0026 UTC. Additionally, the low height of the RFL and the number of IP/RA and FZRA reports prior to this time makes it likely that the transition occurred over a longer period within the domain. In the range-defined QVPs, the low-level $Z_{DR}$ enhancement disappears at approximately 0030 UTC, signaling the disappearance of the RFL in association with a transition to FZRA. We thus estimate that the transition to FZRA occurs at approximately 0030 UTC 4 March.

The Schuur et al. (2012) algorithm is again applied to hourly RAP-derived $T_w$ profiles (Fig. 11). Although the 1900 UTC profile (a 1-h forecast) identifies the emergence of IP at approximately the time when the polarimetric signature first appears, the subsequent RAP analyses are consistent with the IP-to-FZRA changeover time suggested by the observations. The 2100 UTC analysis indicates a $T_w$ profile very near the IP-to-FZRA threshold ($T_{w_{\text{max}}} \approx 1.95^\circ$C; recall FZRA is identified if $T_{w_{\text{max}}} > 2^\circ$C). The 2200 RAP $T_w$ profile (a 1-h forecast) and subsequent times clearly indicate FZRA, suggesting a transition time shortly after 2100 UTC; however, mPING (METAR) reports still indicate IP or IP/RA until 2353 UTC (0033 UTC). Thus, RAP forecast profiles do not capture the transition from IP to FZRA that occurs between 2353 and 0030 UTC. The RAP indicates the transition to FZRA >3 h prematurely, possibly suggesting a warm bias in the RAP model output for this case, or a limitation of the Schuur et al. (2012) algorithm for shallower refreezing heights.

Again, we note the location of the $Z_{DR}$ enhancement in relation to the $-5^\circ$C isopsychrotherm. Prior to 2100 UTC, the $-5^\circ$C isopsychrotherm appears to align itself well with the enhancement in $Z_{DR}$. After 2100 UTC, however, the two diverge because of the possible warm bias of the RAP model. Recall that between 2100 and 0000 UTC, the 2100 UTC cycle of RAP forecasts was used in place of the missing analyses. Once again, the polarimetric radar data and precipitation-type reports
help identify a precipitation transition time in the presence of possible model errors.

4. Synoptic and forecasting considerations

In the three cases examined, the RFL descended in height prior to a transition from IP to FZRA and appeared to intersect the ground at the time of transition. The 3–4 January 2015 case in Albany is investigated further to examine the synoptic setting that led to the transition. The NWS Weather Forecast Office (WFO) in Albany releases radiosondes twice daily at times valid at 0000 and 1200 UTC, allowing for a representative comparison at approximately 0000 UTC when IP were reported in both the METAR and mPING datasets. For the purposes of this discussion, observed soundings are preferred to remove any potential biases or errors in the RAP-analyzed thermodynamic fields.

The 0000 UTC surface analysis provided by the Weather Prediction Center (WPC) shows Albany on the windward side of an inverted surface trough and ahead of an advancing low pressure system (Fig. 12a). The observed sounding (Fig. 12b) reveals a deep, cold near-surface layer with minimum temperatures \( < -7^\circ C \), overlapped by a strong inversion with a maximum temperature of \( 3^\circ C \). The profile is saturated with respect to liquid water throughout much of the lowest 5 km. This environmental setup is different, however, from that found in Nagumo and Fujiyoshi (2017) for a prolonged IP event during which a dry warm layer was present.

Southerly surface winds veer with height to west-southwesterly winds at \( 785 \text{ hPa} \) (Fig. 12b), suggesting low-level WAA via the thermal wind relation. The temperature is \( 0^\circ C \) at this level, indicating the bottom of the \( T > 0^\circ C \) layer. This is consistent with Prater and Borho (1992), who reported that the level at which
winds cease veering with height beneath the melting layer bright band can be used to approximate the location of the lower 0°C temperature height (i.e., bottom of the warm nose) in overrunning FZRA cases. In Prater and Borho (1992), IP were observed when this near-surface cold layer was sufficiently thick. Hanesiak and Stewart (1995) observed that the bottom of the warm nose descends in time during events transitioning from IP to FZRA, which is consistent with the decreasing height of the RFL as observed by the low-level ZDR enhancement in transition events. Note, however, that the bottom of the warm nose (0°C) and the refreezing layer (typically Tw < −5°C) are at different altitudes. Additionally, Hanesiak and Stewart (1995) observed a low-level jet located just above the near-surface cold layer and parallel to the advancing surface warm front. In the present case, however, the low-level jet is aligned more with the surface trough [cf. Figs. 13a and 13b; 31 m s⁻¹ (60 kt) from the south-southwest at 850 hPa].

Surface WAA in the region was weak (Fig. 12c; <5 × 10⁻⁵ K s⁻¹), which allowed low levels to remain <0°C so as to prolong the duration of IP received at the ground. At 850 hPa, however, there was much stronger WAA into the Albany region (Fig. 12d; >5 × 10⁻⁴ K s⁻¹), northeast of an analyzed WAA maximum. The 850-hPa level in this case had a geopotential height of ~1.5 km located a few hundred meters beneath the melting-layer bright band. The approaching WAA maximum was responsible for increasing low-level temperatures. It may also have been responsible for the increasing thickness of the warm layer and, consequently, the RFL descent.

The State College and Wakefield cases presented similar synoptic features with both regions located on the windward side of an inverted surface trough near an approaching surface low pressure system (not shown). Surface WAA in the region was approximately 5 × 10⁻³ K s⁻¹ in both cases, whereas 850-hPa WAA was approximately 10 × 10⁻⁴ K s⁻¹ prior to the transition with an approaching WAA maximum nearby. Again, the descent of the RFL during these transitions from IP to FZRA can be attributed to WAA near 850 hPa increasing the thickness of the warm layer and eroding the near-surface subfreezing layer.

We speculate that the rate at which the RFL descends is related, to first order, to the rate at which the temperature is increasing within the warm nose above the RFL. For example, strong WAA at this level would result in an RFL descending quickly and a transition from IP to FZRA occurring sooner than in a case of weaker WAA. Although the connection between WAA strength and precipitation-type transition timing is intuitive, NWP model errors and biases could make it difficult to diagnose and predict the IP-to-FZRA transition, as demonstrated in the previous section. We suggest that the evolution of the polarimetric refreezing layer ZDR enhancement at this time.

---

6 Interestingly, the height above the surface at which winds start veering is ~916 hPa (820 m AGL), very near the top of the refreezing layer ZDR enhancement at this time.
signature in range-defined QVPs can aid in assessing the changing low-level thermodynamic conditions immediately surrounding the radar prior to a precipitation transition and can help make or refine a forecast of the IP-to-FZRA transition time.

The transition from SN to IP is expected to occur when both melting and refreezing signatures appear in range-defined QVPs. Once the RFL is well defined, tracking its height in time provides insights into changing low-level thermodynamic conditions. For example, the RFL height should remain relatively constant in the event of a prolonged IP event without a precipitation-type transition. As shown above, the RFL descends in time owing to WAA. The cases in section 3 had RFL heights descending at varying rates throughout the event, suggestive of some variation in WAA magnitudes. For an event subject to constant WAA, we speculate that there would be a monotonically and linearly descending RFL from the time of its formation aloft to its disappearance at the ground at the time of the transition to FZRA. Such situations could allow a simple method to forecast the time at which the IP-to-FZRA transition will occur (Fig. 13): assume that the Z_{DR} enhancement associated with the RFL will descend monotonically at a constant rate under continuous low-level WAA. After a sufficient number of volume scans show the RFL descent, one may extrapolate a line through the middle of the enhancement forward in time until it intersects the ground, indicating the approximate transition time. Although this method is only appropriate for the small domain near an operational WSR-88D radar and not over an entire forecasting area, valuable insight may still be gained that could help make informed decisions about the rest of the area.

Though crude, this method provides a quick way to assess current trends and adjust forecasts accordingly. For example, a quickly descending RFL could indicate an imminent transition to FZRA, whereas a steady or slowly descending RFL indicates a delayed transition time and prolonged IP. Alternatively, we speculate that an ascending RFL and/or descending melting layer may indicate a potential transition to SN, the most prevalent precipitation type to follow IP in the case of a narrowing melting region with the transition from the warm to colder air masses on the backside of winter storm systems (Cortinas et al. 2004). Regardless of the transition type, WSR-88D radar data update from approximately every few to tens of minutes, allowing for the continual monitoring of low-level conditions in real time to refine short-term forecasts as needed. In contrast, high-resolution NWP models such as the RAP update hourly and are subject to inherent biases and errors.

5. Conclusions

In conjunction with surface precipitation reports and RAP model output, we analyzed the evolution of the polarimetric refreezing signature during precipitation transition events for the first time. We developed an extension of the quasi-vertical profile (QVP) approach and applied it to ice pellet (IP) to freezing rain (FZRA) transition events. These so-called range-defined QVPs allow the evolution of polarimetric signatures to be visualized with good resolution at all levels. By using data from all available elevation angles at each height, all low- to midlevel features such as the refreezing layer (RFL), melting layer, and planar crystal growth zones are well resolved. This is an improvement over previous techniques that utilize only one elevation angle.

The evolution of range-defined QVPs in three IP-to-FZRA events reveals a repeatable trend of RFL descent over time leading up to the precipitation-type transition. The time at which the RFL appears to intersect the ground indicates the approximate time that surface precipitation reports transition from IP to FZRA. Precipitation reports from crowdsourced mPING and automated and/or augmented ASOS sources were used to determine the likely precipitation types and identify the precipitation-type transition times. The combination of data from both sources mitigates the biases associated with each individual data type so as to provide a comprehensive view of all potential precipitation types that may have occurred during each of the three events. The consistency of both precipitation-type reports with range-defined QVP polarimetric signatures in each event (e.g., IP reports in association with both an ML and RFL) makes this type of analysis a viable approach for matching surface-based precipitation types with polarimetric observations. Though this study only considered three cases, these preliminary results are encouraging. Future work involving more cases will assess the reliability of range-defined QVPs and surface precipitation types for winter-precipitation-type transition events.

The utility of range-defined QVPs and surface precipitation-type identification was explored by comparing the analysis to the Schuur et al. (2012) precipitation classification algorithm based on RAP model-derived T_w profiles. Although the algorithm indicated that the RAP analyses performed well in section 3a, biases in the model were evident with a warm nose in section 3b that was too shallow and did not capture an IP-to-FZRA transition, and in section 3c with a premature IP-to-FZRA transition time. Although the RAP-based algorithm had varying success in depicting an IP-to-FZRA transition in each event,
the range-defined QVPs and surface-based precipitation types all consistently depicted a transitioning winter precipitation event. By comparing real-time polarimetric and precipitation-type observations with NWP model analyses, the reliance on such models can be adjusted for short-term forecasts.

The synoptic conditions favorable for an IP-to-FZRA transition event were examined. These events are distinct from prolonged IP events in which IP conditions are produced via strong evaporative cooling of rain falling through a cool, dry layer (e.g., Kumjian and Fujiyoshi 2017). Here, IP form when frozen hydrometeors melt within an elevated warm layer and refreeze within a near-surface \( T_w < 0^\circ \text{C} \) layer. FZRA in the cases presented herein is produced when the surface \( T_w < 0^\circ \text{C} \) layer is insufficient in depth or strength to fully refreeze hydrometeors prior to contact with the surface. Moderate WAA located within the warm nose warsms and deepens the layer, which is believed to be responsible for the descent of the RFL in such IP-to-FZRA transition cases. It is also suggested that the descent of the RFL is related to the rate of low-level WAA. In cases of constant WAA, we propose that extrapolation of the RFL trend in range-defined QVPs can be used to forecast the time of the transition from IP to FZRA.

At present, the National Weather Service does not display polarimetric radar data in QVPs. We suggest that this method of analysis be implemented in the NWS Advanced Weather Interactive Processing System (AWIPS) for easy RFL monitoring. Though the ingestion of all volume scan data to create range-defined QVPs can be computationally expensive in an operational setting, this and other modified QVP techniques seem promising for monitoring short-term trends in winter precipitation and anticipating precipitation-type transitions.

**Acknowledgments.** The authors acknowledge funding from NSF Grant AGS-1143948. We thank the anonymous reviewers of this and a previous version of this manuscript whose comments are invaluable to the development of the current methodology and focus of this paper. We also thank Robert Schrom (Penn State) for assistance with reading in RAP model data and Sonia Lasher-Trapp (University of Illinois) for a helpful discussion. ASOS data were obtained from the Iowa Environmental Mesonet website. RAP and radar data were obtained from the National Centers for Environmental Information website. This study is part of DMT’s M.S. thesis and current Ph.D. research at Penn State.

**REFERENCES**


