Polarimetric Radar Observations in the Ice Region of Precipitating Clouds at C-Band and X-Band Radar Frequencies

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ABSTRACT

Data collected by C-band and X-band radars in northwestern Italy are analyzed to study the behavior of the polarimetric variables in the ice region of precipitating clouds, with special emphasis on the specific differential phase $K_{dp}$. It is found that stratiform precipitation, irrespective of the precipitation type at the ground and as opposed to convective systems, is characterized by well-pronounced positive differential reflectivity $Z_{dr}$ and $K_{dp}$ values near the model-predicted $-15^\circ C$ isotherm. The regions of enhanced $Z_{dr}$ and $K_{dp}$ are likely related to the growth of dendrite crystals in the region where the difference between the saturation vapor pressure over water and the saturation vapor pressure over ice is greatest. Coincident C-band and X-band measurements, in conjunction with electromagnetic scattering simulations, demonstrate that $K_{dp}$ scales with frequency, indicating that the ice particles in the vapor deposition preferential growth zone are Rayleigh scatterers. Peak values around 2.0° and 3.5° km$^{-1}$ are observed at C band and X band, respectively. Most noteworthy is that an extended analysis of hourly and daily vertical profiles of C-band data over 1 year has shown that $K_{dp}$ observations around the $-15^\circ C$ temperature level in stratiform precipitation are well correlated (0.8) with the reflectivity in the underlying rain layer.

1. Introduction

During the last couple of decades several studies in the literature have focused on observations of ice particles with polarimetric radars. These studies often included the analysis of in situ aircraft measurements and mainly considered the radar observations of differential reflectivity (Bader et al. 1987; Wolde and Vail 2001a,b; Hogan et al. 2002) and linear depolarization ratio (Matrosov et al. 1996, 2001; Wolde and Vail 2001a,b).

Several papers considered specific differential phase shift $K_{dp}$ measurements of ice particles. The quantity $K_{dp}$ is defined as one-half of the range derivative of the two-way propagation differential phase $\Phi_{dp}$, which is the phase shift occurring between the horizontally and vertically polarized pulses along the propagation path (Seliga and Bringi 1978). Some of these studies attributed the $K_{dp}$ signatures in the ice region of convective storms to the effects of cloud electrification (Hendry and McCormick 1976; Caylor and Chandrasekar 1996; Carey and Rutledge 1996; Ryzhkov and Zrnić 2007). Only a few studies have reported observations of differential phase in snow (Hendry et al. 1976; Vivekanandan et al. 1994; Ryzhkov and Zrnić 1998; Trapp et al. 2001; Kennedy and Rutledge 2011). For example, Hendry et al. (1976) reported values of $K_{dp} \sim 0.4^\circ$ km$^{-1}$ in snow at S band and values up to $1^\circ$ km$^{-1}$ at Ku band in heavy snow. Vivekanandan et al. (1994) have also shown S-band differential phase shift changes by $10^\circ$ above the bright band in the stratiform region of a mesoscale convective system, with peak values of $K_{dp}$ up to $0.5^\circ$ km$^{-1}$.
More recently, positive values of $K_{dp}$ in the ice region of stratiform precipitating clouds have been shown to be linked to increased surface precipitation rates at S band (Kennedy and Rutledge 2011). These authors found regions of enhanced $K_{dp}$ are observed near the $-15^\circ$C isotherm and suggested that they are related to the growth of dendritic crystals. The observed values of $K_{dp}$ are only few tenths of a degree per kilometer at S band, but are expected to be higher and more easily detectable at higher frequencies such as those of C and X band because of the $K_{dp}$ scaling with frequency (Bringi and Chandrasekar 2001, their Eq. (7.101)). Elevation scans at high space resolution collected by the Arpa Piemonte transportable X-band radar system (ARX; Bechini et al. 2008) during summer stratiform precipitation, for example, show $\Phi_{dp}$ increases in excess of 40° in the ice region, with $K_{dp}$ peak values up to $-2^\circ$ km$^{-1}$ in the midtropospheric layer at 5–7-km height (Fig. 1), corresponding to temperatures in the range from $-9^\circ$ to $-20^\circ$C as measured by a nearby radio sounding.

To augment the work of Kennedy and Rutledge (2011), which focused on S-band observations for selected winter snowfall case studies, this study aims at characterizing the phenomenological behavior of $K_{dp}$ in the ice region of generic mesoscale precipitation systems, irrespective of the phase of precipitation (liquid or solid) at the surface, at C-band and X-band radar frequencies.

Differential phase shift measurements have the well-known advantages of being 1) immune to radar calibration, 2) immune to attenuation due to propagation, and 3) mainly insensitive to partial beam blocking (Zmič and Ryzhkov 1999; Bringi and Chandrasekar 2001; Giangrande and Ryzhkov 2005; Ryzhkov et al. 2005; Friedrich et al. 2007). One of the potential problems with differential phase measurements is the backscattering differential phase $\delta$ at C and X band. In fact radar does not measure directly $\Phi_{dp}$, but the total differential phase shift $\Psi_{dp}$, which is equal to the sum of the differential propagation phase and $\delta$, that is, $\Psi_{dp} = \Phi_{dp} + \delta$. Typical ice particles in the ice region of stratiform clouds around $-15^\circ$C are Rayleigh scatterers at both C-band and X-band radar frequencies (section 3). This fact makes $\delta$ negligible (Jameson and Mueller 1985), removing a potential source of error in deriving reliable specific differential phase estimates.

The primary source of data for this paper is the C-band Bric della Croce radar, close to Turin (Italy). The data from this system allowed an extensive analysis of vertical profiles of the polarimetric measurements in the northwest Italy subalpine region. Although $K_{dp}$ is the main focus of this study, differential reflectivity $Z_{dr}$ measurements from this radar are exploited to infer the dominant crystal habit. Additionally, measurements collected by a transportable X-band system during a selected event are also considered. This made possible a thorough analysis of $K_{dp}$ in the ice region with simultaneous observations at C-band and X-band radar frequencies (5.640 and 9.375 GHz, respectively). Interpretation of radar measurements at the $-15^\circ$C isotherm level are supported by electromagnetic scattering simulations performed at these frequencies.

**FIG. 1.** Elevation scan from Col de Tende at 1743 UTC 14 Aug 2010, along the 340° azimuth, showing (a) reflectivity and (b) filtered differential phase shift. The underlying topography is shown in gray. The $K_{dp}$ (defined as one-half the range derivative of $\Phi_{dp}$) contours are overplotted in both panels: isolines at 0.6° (dotted line), 1.1° (solid line), and 1.6° km$^{-1}$ (thick solid line). The horizontal gray lines mark the location of the 0° and $-15^\circ$C temperature levels as inferred from the nearby radiosounding of Cuneo Levaldigi (World Meteorological Organization code 16113).
The paper is organized as follows. Section 2 describes the radar characteristics and data processing. In section 3, we present a widespread rainfall case using observations made simultaneously with C- and X-band radar systems. These coincident multifrequency observations are used to study the polarimetric signatures in the ice region of clouds, together with electromagnetic scattering simulations of ice-particle distributions. In section 4, the main findings about the $K_{dp}$ signatures in the ice region from the single case analysis (section 3) are confirmed and further extended by means of a statistical analysis over more than 1 yr of radar vertical profiles from the operational C-band radar. The main conclusions are discussed in section 5.

2. Radar data and processing

Arpa Piemonte (Environmental Protection Agency of Piedmont) operates a dual-polarization radar network composed of two operational C-band systems and one research transportable X-band radar. For this work the measurements collected by the ARX radar and from one of the C-band systems, the Bric della Croce radar (Bric), are considered.

The Bric radar is located at 736 m MSL on the top of the Turin hill. The X band was deployed since 2008 at different locations for specific measurement campaigns. In 2009 (data analyzed in section 3) the radar was deployed in Carmagnola, 235 m MSL, 16 km south of Turin, where a collocated rain gauge and a Thies-Clima optical laser disdrometer are also available. During the summer of 2010 the transportable radar was deployed on the mountain pass of Col de Tende (1835 m MSL), on the border between Italy and France, for a measurement campaign within the Italian–French cooperative project known as “Gestion des Crues par Integration des Systemes Transfrontaliers de Prevision et de Prevention des Bassins Versants Alpins” (CRISTAL; Cremonini et al. 2010). The main technical characteristics of the systems are listed in Table 1 and a map of the region with radar deployment locations is shown in Fig. 2. The X-band volume coverage pattern includes a plan position indicator (PPI) at vertical incidence for vertical Doppler observations and differential reflectivity calibration.

To obtain the moment data used for this paper Doppler-based filtering was not applied, relying on a postprocessing fuzzy logic scheme adapted from Liu and Chandrasekar (2000) to classify meteorological and nonmeteorological echoes. A distinctive feature of this implementation is that clutter (all nonmeteorological echoes) is treated exactly as an additional hydrometeor type and is identified within the same fuzzy logic volume processing, allowing a reduced total computation time (Davini et al. 2012). Giuli et al. (1991) exploited the properties of radar measurements to distinguish between meteorological and nonmeteorological echoes. Their approach was based on a number of subsequent tests, specifically including the Doppler velocity (low magnitude for clutter echoes), the spatial variability of $Z_{dr}$ (high variance for clutter), and a clutter map test (difference between the actual and a statistical clear-sky reflectivity map). Following this approach we defined probability functions (instead of fixed thresholds) for inclusion in the fuzzy logic scheme, and we additionally considered the correlation coefficient $\rho_{HV}$. Lower values in clutter ($Z_{dr}$), and the spatial variance of $\Phi_{dp}$. This algorithm is operationally implemented on the Arpa Piemonte radar network, producing real-time maps of the radar echo type—clutter, large drops, drizzle, rain, heavy rain, hail mixed with rain, hail, graupel, wet snow, dry snow, or crystals. However, for the purpose of this study we are only interested in the identification of the clutter-contaminated volume bins. An asymmetric beta distribution is used as the form for the radar membership functions, to allow more flexibility in shaping the probability density. After the clutter bins are flagged in the polar volumes, $\Phi_{dp}$ is filtered using the Hubbert and Bringi (1995) scheme. Then the rain profiling algorithm based on Testud et al. (2000) is applied to correct the reflectivity for path attenuation.

<table>
<thead>
<tr>
<th>TABLE 1. Main characteristics of the C-band (Bric della Croce) and X-band (ARX) radar systems. MDS is minimum detectable signal, and HV is horizontal and vertical polarization.</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>C-band Bric della Croce</strong></td>
</tr>
<tr>
<td>Antenna diameter (m)</td>
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<tr>
<td>Antenna beamwidth (°)</td>
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<td>Antenna gain (dB)</td>
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<td>Polarization type</td>
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<td>Transmitter peak power (kW)</td>
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<td>Range resolution (m)</td>
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<td>Receiver dynamic range (dB)</td>
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<td>MDS (dB)</td>
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<tr>
<td>Volume scan strategy</td>
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<tr>
<td>Range (km)</td>
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<tr>
<td>No. of sweeps</td>
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<tr>
<td>Elevations (°)</td>
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<tr>
<td>Pulse length (μs)</td>
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<tr>
<td>PRF (Hz)</td>
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<tr>
<td>Sensitivity at 50 km (dBZ)</td>
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<td>Scan frequency (min)</td>
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Attenuation correction is performed exclusively in the rain medium (i.e., below the melting level) to avoid introducing a bias due to differential phase increases in the ice region.

The $K_{dp}$ is estimated using the Wang and Chandrasekar (2009) technique. This technique has been implemented in the operational postprocessing chain of the Arpa radar network and proven to be suitable for the unsupervised processing of large datasets. In fact it does not require the estimation of the system differential phase (which may typically fluctuate with time as a consequence of hardware maintenance) and is insensitive to differential phase aliasing. The $K_{dp}$ estimation requires measurements with a good signal-to-noise ratio (SNR), to avoid too-noisy phase measurements. A 5-dB threshold was used for this study, which implies that the effective sensitivity of the two radar volume scans using the short pulse (Table 1) is decreased from 2 dBZ (4 dBZ) at 50-km range to 7 dBZ (9 dBZ) for the C-band (X band) radar, respectively.

The vertical profiles of temperature needed for the freezing-level identification in the attenuation correction procedure and for the $K_{dp}$ analysis on which this paper is focused, are taken from the 3-hourly output of the local area Consortium for Small-scale Modeling (COSMO) model (http://www.cosmo-model.org) run over Italy (analysis at 0000 UTC and subsequent 3-hourly forecasts). In previous verification studies (COSMO Newsletter 7, chapter 5.4, available online at http://www.cosmo-model.org/content/model/documentation/newsLetters/newsLetter07/cnl7_schubiger.pdf) the accuracy of midtropospheric temperature analysis and 24-h forecast has been shown to be within 1°C.

The vertical profiles are calculated considering all the radar volumes collected with a 5-min frequency in a given time interval (typically 1 or 24 h). Only sweeps at elevation angles below 15° are used since the polarimetric signals decrease with higher view angles because of the general hydrometeor orientation with the symmetry axis in the vertical direction. Additionally, range bins farther than 50 km from the radar are excluded to avoid excessively sparse vertical sampling. The data are then binned in 0.3-km height levels. The lowest available height is 900 m (layer 750–1050 m) for the C-band radar and 300 m (layer 150–450 m) for the X-band radar.

Assessment of the $Z_{dr}$ calibration and impact of differential attenuation

Differential reflectivity measurements from the C-band radar are used in the following sections, in addition to reflectivity $Z_h$, $K_{dp}$, and $r_{HV}$, to support the characterization of the dominant crystal habit limitedly to stratiform precipitation systems. As shown in Wolde and Vali (2001a) and more recently in Williams et al. (2011), strongly distinctive $Z_{dr}$ signatures are associated with either dendrite crystals or plate crystals. Plate crystals, owing to their high volume fraction of ice and low axial ratio may give $Z_{dr}$ values up to 6–7 dB, while $Z_{dr}$ for dendrites with lower bulk density attains values in the range 1–2 dB. The required accuracy on the $Z_{dr}$ measurements to distinguish between the aforementioned crystal habits is therefore not as stringent (say 0.1–0.2 dB) as for use in quantitative polarimetric rainfall estimation algorithms.

Figure 3 shows a scatterplot of the hourly average $Z_{dr}$ versus $Z_h$ for the lowest level (900 m) of 260 vertical hourly profiles in stratiform precipitation (refer to section 4 for the profiles processing and the stratiform/convective classification). Only profiles with the freezing level above 1.5 km are considered to ensure the measurements are in the rain medium. The $Z_{dr}$ values tend to approximately 0 dB when the reflectivity is below 20 dBZ, as expected because of the mostly spherical shape of small droplets. Displayed for comparison (dashed line) is also the $Z_{dr}$ versus $Z_h$ average relation from Bringi et al. (2001).
region, where the radar pulse is sensing the cloud after passing through the rain and the melting layer. To check the effect on the average profiles we derived statistics from large dry aggregates right above the melting layer in stratiform precipitation. Low-density dry aggregates are known for the intrinsic low $Z_{\text{dr}}$ (Ryzhkov et al. 2005). The vertical profiles of $Z_{\text{dr}}$ show a local minimum just above the freezing level (Fig. 14) where large aggregates are most likely to be found. The temperature corresponding to this minimum for the daily profiles in stratiform precipitation (27 events between April 2009 and December 2010; section 4) is $22.6 \pm 1 ^{\circ} \text{C}$ and the associated average $Z_{\text{dr}}$ is $0.23 \pm 0.14$ dB, varying in the range from $-0.06$ to $0.52$ dB. Although these values are slightly higher than the 0.25-dB upper limit reported in Ryzhkov and Zrnić (1998) for cold snow, the limited variability over the long period analyzed is a good indication of the stability of the system and ensures the considered profiles are not significantly affected by differential attenuation. Overall, the $Z_{\text{dr}}$ average profiles for the stratiform precipitation cases can be considered as accurate within 0.2 dB, which is enough for the purpose of crystal habit identification.

3. Widespread stratiform precipitation case

Dual-polarization radar observations at C band (Bric della Croce) and X band (ARX) during an intense widespread rainy event are analyzed in this section. The event occurred on 27 April 2009 and was chosen because it can be considered as representative of the typical widespread precipitation events interesting northern Italy during autumn and spring. Southerly moist flow associated with the passage of a warm front helped produce continuous rain for most of the day with 24-h-average accumulation of 76 mm over the plains, and peak values up to 140 mm. In the northern mountains, values above 200 mm were recorded by the regional tipping-bucket rain gauge network. The study area for the following analysis is the union of the 50-km-range domains of the two radars, spaced only 16.8 km apart (Fig. 2). This area encompasses mainly the plains of the Po River valley and includes the foothills of the western and northern Alps.

The X-band radar performed a scan strategy that included an acquisition at vertical incidence within each 5-min volume scan. Doppler velocity observations at vertical incidence (Fig. 4) allow to infer that riming was likely not relevant in the vicinity of the radar. In fact, the lack of a significant number of particles with Doppler velocity lower than $-1.5$ m s$^{-1}$ (downward) is here taken as an evidence of the negligible generation of graupel by heavy riming (Zawadzki et al. 2001). However, the capture of small supercooled cloud droplets by snow crystals can still take place when the air is saturated over water and simultaneous growth of ice particles and liquid droplets is effective (Korolev and Mazin 2003; Korolev 2007) without leading to the formation of heavily rimed crystals like graupel.

One possible way to extend the above inference to a larger area surrounding the radar is by considering the probability distributions of the 1-h rainfall recorded by the 90 rain gauges in the area. Figure 5 shows the distribution of a sample including 1-h rainfall amounts from a rain gauge collocated with the radar ($-30$ m) and a second distribution including all other rain gauges in the study area (89 gauges). The nonparametric Kolmogorov–Smirnov test over the two samples has a $p$ value $= 0.60$, leading one to accept the null hypothesis that the samples are drawn from the same distribution. The similarity of the two distributions makes it possible to extend to the whole study area the Doppler-based inference that graupel is likely not relevant for this event. The precipitation largely originates from microphysical processes characteristic of stratiform rains, although it is not possible to completely exclude a minor role of local convection. The ARX vertical observations are therefore considered as suitable for investigating ice-particle distributions mainly resulting from vapor deposition, light riming, and aggregation.
a. Polarimetric radar observations

Figure 6 shows the C-band polarimetric radar measurements collected at 0025 UTC during a PPI scan at the elevation of 7.4°. Both $Z_{\text{H}}$ and $Z_{\text{DR}}$ show the characteristic increase in the melting layer, with a corresponding low correlation coefficient $\rho_{\text{HV}}$ in the range 0.8–0.96, deriving from the coexistence of liquid and partially frozen particles (Ryzhkov and Zrnić 1998). The values of $\rho_{\text{HV}}$ in the western portion of the domain are in general lower because of the weaker radar echo (low SNR), and the $Z_{\text{DR}}$ field is consequently noisier. At higher levels, on the east side of the radar domain, $Z_{\text{DR}}$ shows marked positive values (up to 2.2 dB) along an azimuth sector about 135° wide centered on the $-15^{\circ}$C temperature level at ~4.7 km MSL. The corresponding reflectivity varies in the range 10–25 dBZ. These values are fairly consistent with the “category A” microphysical regime (dendrite crystals) described in Williams et al. (2011), who found the enhanced $Z_{\text{DR}}$ layer (1–3 dB) in different case studies at temperatures between $-13^{\circ}$ and $-15^{\circ}$C, and with corresponding reflectivity between 10 and 30 dBZ. Similar $Z_{\text{DR}}$ signatures in this temperature range were also previously reported by Sauvageot et al. (1986), Wolde and Vali (2001a,b), Andric et al. (2009), and Kennedy and Rutledge (2011). In particular Wolde and Vali (2001a) analyzed in situ measurements with microphysical probes and W-band airborne polarimetric observations, reporting $Z_{\text{DR}}$ up to 2 dB for dendritic crystals within a nimbostratus cloud at $-13^{\circ}$C.

The $K_{\text{DP}}$ field (Fig. 6c) shows a similar behavior as $Z_{\text{DR}}$ at this time, with the highest positive values around 1.3° km$^{-1}$ at $X = 20$ km, $Y = -25$ km. It is interesting to note that the positive signatures ($Z_{\text{DR}}$ and $K_{\text{DP}}$) lie in a region of relatively low correlation coefficient $\rho_{\text{HV}}$ (Fig. 6d). The aforementioned positive peak appears confined in an area where $\rho_{\text{HV}} < 0.98$, with higher correlation coefficient below (closer ranges). The same is observed on the tongue-shaped positive $Z_{\text{DR}}$ signature extending to the north, around $X = 28$ km, $Y = 5$ km. In general, the region where $Z_{\text{DR}} \geq 1$ dB at altitudes where the temperature is lower than $-5^{\circ}$C is characterized by $\rho_{\text{HV}}$ in the range 0.95–0.98. The transition from relatively low correlation and high $Z_{\text{DR}}$ to higher correlation

[Figure 4. Frequency plot (gray scale is the number of observations $N_{\text{obs}}$ in logarithmic units, using 0.3 m s$^{-1}$ and 0.125-km intervals for velocity and height, respectively) of daily Doppler velocity observations from vertical-looking X-band radar scans (27 Apr 2009). The horizontal gray lines mark the temperature levels of 0° and $-15^{\circ}$C, while the vertical gray line marks the 0 m s$^{-1}$ velocity.]
of ''patches'' of peak only increasing from 2.2 to 2.4 dB. The appearance of larger aggregates (higher reflectivity above the freezing level) that require more time to melt. On the other hand the main positive drift values aloft did not change significantly between 0025 (Fig. 6) and 0155 UTC (Fig. 7), the Kdp field shows a substantial increase, from a peak value of 1.3 m−1 at 0025 UTC to 2.3 m−1 at 0155 UTC (southern sector). In addition, unlike the rather good spatial matching between the Zdr and Kdp positive patterns at 0025 UTC, 90 min later we note how Kdp extends to lower heights, particularly in the eastern sector (Kdp > 0.5 m−1 down to the −5°C temperature level) in a region of low differential reflectivity and high \( \rho_{HV} \) (>0.98). While Zdr is only related to the (reflectivity weighted) axis ratio and density of the hydrometeors, Kdp is also a function of the mass of the crystals [Bringer and Chandrasekar 2001, their Eq. (7.101)]. So when aggregation starts to reshape the particle size distribution (PSD) near the −15°C level (see also comments about Fig. 16 in section 4) both the axis ratio and density of the largest particles are reduced, likely causing the sudden decrease of the Zdr values, but a rather smoother decrease of Kdp. This behavior will be shown to be quite a common feature of the vertical profiles of polarimetric measurements in stratiform precipitation (section 4).

To compare the radar observations at C band and X band in the ice region, we computed the average values in a 1.3-km-depth layer (4.2–5.5 km MSL) centered on the −15°C temperature level. Figure 8 shows the resulting average Kdp fields over the study area at 0155 UTC. There is excellent agreement between the two radar estimates, partly owing to the proximity of the two systems. The X-band Kdp estimates are higher resolution relative to the C-band estimates because of the difference in range resolution (125 vs 340 m; Table 1).

Hourly average vertical profiles of the polarimetric radar measurements have been calculated for both radars. In addition to the common statistical parameters (mean and standard deviation), the 20 quantiles (vigintiles) have also been calculated (Jain and Chlamtac 1985) to better represent the distribution of radar measurements, which are in many instances highly non-Gaussian. In fact, the quantiles divide the sample in 20
subsets (5% spaced classes) containing approximately the same number of observations and represent the boundaries between the resulting equal-populated classes. This allows a more comprehensive description of the data and an effective graphical representation of the distribution, with special emphasis on extreme values. Conversely, a simple average of hourly or daily profile may hide significant polarimetric signatures resulting from a transient microphysical process.

Figures 9 and 10 show the \( Z_h \) and \( K_{dp} \) quantiles distribution with height respectively for the C-band and the X-band radar data, corresponding to 1 h (0400–0500 UTC) of light to moderate precipitation. The quantiles are only plotted between 10% and 90% to exclude possible outliers from the visualization. The overlaid average \( K_{dp} \) profiles at the two frequency bands show a peak between 4.5 and 5.0 km MSL, roughly corresponding to the temperature interval from \( -13^\circ C \) to \( -15^\circ C \) as per the 0300 and 0600 UTC COSMO model forecast.

In general, \( K_{dp} \) in the ice region at X band is observed to be about 1.7 times \( K_{dp} \) at C band, in good agreement...
with the ratio of the operating frequencies of the two radars. Peak values of observed $K_{dp}$ during the event are $\sim 2^\circ$ km$^{-1}$ at C band and $\sim 3.5^\circ$ km$^{-1}$ at X band, respectively. Such values are consistent with the scattering simulations reported in Kennedy and Rutledge (2011). When $K_{dp}$ values are well above its typical measurement error (i.e., $K_{dp} > 0.2^\circ$ km$^{-1}$ at C band), $K_{dp}$ fields at the $-15^\circ$C level at C and X band are highly correlated (the Pearson’s correlation for the pairs of $K_{dp}$ maps is $r > 0.9$), indicating the robustness of the $K_{dp}$ estimates in the ice region. Reflectivity fields instead are less correlated ($0.7 < r < 0.9$), likely because of the variability of the response of C versus X band to a given PSD and to path attenuation uncertainties determined both in rain and melting layer.

Figure 11 shows the frequency plot of $K_{dp}$ at C and X band at the $-15^\circ$C level for the first 10 h of the day, when the highest rainfall amounts were recorded by the surface gauges in the area (accumulations between 12 and 51 mm). The relation between the two frequencies appears roughly linear, although for higher $K_{dp}$, the C-band values underestimate the corresponding X-band scaled estimates. This behavior only concerns the peak values and it is believed to be a simple consequence of the different range resolution (Table 1), for which the higher range resolution at X band allows us to detect more intense local maxima.
b. Comparison with electromagnetic scattering simulations

T-matrix simulations of scattering at C and X band have been performed to interpret radar measurements at the $-15^\circ$C isotherm level, using simplified assumptions based on those adopted by Kennedy and Rutledge (2011). The population of particles at the $-15^\circ$C level is modeled as a mix of highly oblate spheroid crystals for diameter smaller than 3 mm with axis ratio varying between 0.05 and 0.15 and bulk density values as reported in Heymsfield et al. (2004) and aggregates for larger diameters. Exponential PSD are assumed with $N_0$ varying between $50 \times 10^3$ and $400 \times 10^3$ cm$^{-3}$ and $\Lambda$ between 25 and 45 cm$^{-1}$ (Lo and Passarelli 1982; Kennedy and Rutledge 2011). Simulations show that...

![Diagram](image1)

**Fig. 8.** Average $K_{dp}$ in the layer 4.2–5.5 km MSL ($-15^\circ$C), at 0155 UTC 27 Apr 2009, for (a) C-band radar and (b) X-band radar. A 50-km-range ring is overplotted for the C-band radar (solid line) and the X-band radar (dashed line) in both panels. For ease of comparison, the color palettes are scaled by 1.67, corresponding to the ratio of the operating frequencies.

![Diagram](image2)

**Fig. 9.** The distribution (a) of reflectivity and (b) $K_{dp}$ at all height levels ($y$ axis, 0.3-km vertical spacing) for the 0400–0500 UTC data collected on 27 Apr 2009 by the C-band radar. Gray shades represent the values of the quantiles ($x$ axis). The black line in both panels represents the average profile (top $x$ axis), and the dashed line marks the height of the $-15^\circ$C level.
$Z_h$ values are not significantly influenced by the frequency, $K_{dp}$ can be scaled according to the ratio of wavelengths, and finally, resonance effects such as the differential phase shift upon backscattering are negligible at both frequencies.

Figure 12 shows the observed $K_{dp}$ versus $Z_h$ dispersion at C band (Fig. 12a) and X band (Fig. 12b), together with the results from the scattering simulations. The observed radar variables are the averages in the 4.2–5.5-km-height layer during 10 consecutive hours, as in Fig. 11. Given the influence of wet radome attenuation on X-band power measurements, the reflectivity values for each scan have been corrected based on the 1-min rain-rate measurements from the collocated optical disdrometer (Bechini et al. 2010). A fairly good agreement at C band is noted for the positive $K_{dp}$ values. The agreement is slightly worse at X band because of the higher dispersion of the observations and a reflectivity bias of about $-1.4$ dB, relative to the C-band measurements. The higher dispersion of the X-band reflectivity is likely related to the increased uncertainty affecting the power measurements due to 1) residual wet radome attenuation and 2) underestimation of the path attenuation. The average $-1.4$ dB bias between the X-band and C-band radar may also be a result of the excess attenuation at the higher operating frequency.

4. Analysis of polarimetric vertical profiles from C-band radar observations

Space–time average vertical profiles allow a statistical evaluation of the vertical structure of the atmosphere from radar measurements. Specifically the relation between measurements at different height levels can be addressed without explicitly considering the height-dependent advection, relying on the assumption that within a given observation period (~hours) the observed cloud water content aloft will mainly precipitate within the same (sufficiently large) sampling area.

More than 1 yr of data routinely collected at C band between 2009 and 2010 by the operational radar was analyzed leading to a selection of 54 significant rainy days. This choice was objectively made based on the regional rain gauge network and selecting only the days
when at least 5 mm of cumulative precipitation was recorded by at least one rain gauge over the area of interest, defined in this instance by the 50-km-range area around the C-band radar only.

Daily and hourly profiles were calculated for all polarimetric measurements, allowing us to work on a wide statistical sample and check possible differences at the two time scales. To perform an objective partitioning of the dataset into stratiform/convective cases, the characteristic distribution of reflectivity vertical profiles described for example in Steiner et al. (1995) and Yuter and Houze (1995) is exploited to define a metric for convection named radar convective parameter (RCP), described in the appendix. For the purpose of the following analyses we arbitrarily define an event to be stratiform when RCP is lower than 50th percentile and convective elsewhere. Figure 13 shows the histograms of the resulting monthly distribution of stratiform and convective days. A well-defined seasonal distribution arises from the automated classification, with convective events mainly concentrated in the summer months between June and August and stratiform events more...
frequent in autumn and spring. May is typically a transition period, with both stratiform and convective events. Winter is the driest season in northern Italy, and the dataset analyzed is not an exception, with no significant precipitation between January and March.

a. Polarimetric vertical profiles and their relation with crystal habit

Figure 14 shows the hourly vertical profiles of the four radar variables $Z_h$, $Z_{dr}$, $K_{dp}$, and $\rho_{HV}$, with colors representing the RCP value. The decreasing magnitude of the correlation coefficient (Fig. 14d) for lower RCP values can be ascribed to the bias introduced by the conventional zero-lag $\rho_{HV}$ estimator at low SNR (Lei et al. 2012), more likely to occur in stratiform precipitation. The change from stratiform to convective conditions can be clearly seen by the vanishing of the brightband nose in the reflectivity profiles, fairly correlated with the increasing RCP. The melting layer causes a well-defined local enhancement $Z_h$, $Z_{dr}$, and $K_{dp}$, occurring (with slight differences, depending on the considered measurements) just below the freezing level, evident in profiles with RCP lower than approximately 5 dB (see also the solid thick lines in Fig. 14, representing the average of all daily profiles). The melting layer is also very well captured in the profile of $\rho_{HV}$, by virtue of its sensitivity to the diversity of orientation, shape, size, and thermodynamic phase of the hydrometeors. Interestingly, the melting-layer signature is well defined in $Z_{dr}$ and $\rho_{HV}$ also for profiles with higher RCP, that is, with increasing convective characteristics (Baldini and Gorgucci 2006). The increase around the $-15^\circ$C level on the other hand is only depicted in the $Z_{dr}$ and $K_{dp}$ profiles. As already noted in section 3, the peak

![Figure 14. Hourly vertical profiles of C-band (a) $Z_{dr}$, (b) $Z_{dr}$, (c) $K_{dp}$, and (d) $\rho_{HV}$ colored according to their respective RCP values. The RCP quantiles (0%, 25%, 50%, 75%, 100%) are, respectively, 1.1, 2.7, 3.9, 7.3, and 21.7 dB. The black (gray) thick lines represent the average of the daily profiles for stratiform (convective) events. It is possible to note several $Z_{dr}$ convective profiles (high RCP) notably affected by differential attenuation (negative values up to $-1$ dB). To highlight the variations for small values, the $K_{dp}$ profiles are plotted on a log axis.](image-url)
appears sharper in $Z_{\text{dr}}$ than in $K_{\text{dp}}$, likely as a consequence of the different shape and mass dependency of the two variables. The analysis of $Z_{\text{dr}}$ reveals that the peak around $-15^\circ\text{C}$ in the hourly profiles for stratiform precipitation varies between 0.1 and 1.3 dB (mean values), while the 90th percentile span over 0.5–2.5 dB (Fig. 15), without any apparent relation with the reflectivity. Instead, $K_{\text{dp}}$ shows (gray shades and size of the filled circles in Fig. 15) a fairly defined increase with reflectivity, which is further discussed in the next subsection. Also noteworthy is the sudden increase in $Z_{\text{dr}}$ peak encountered in this transition to a warmer layer, where aggregation produces larger particles with lower density and the overall anisotropy of the medium is reduced.

The generally moderate $Z_{\text{dr}}$ peak encountered in this study ($Z_{\text{dr}} \leq 2.5$ dB for 90% of the data) suggests that values as high (6–7 dB) as reported by Wolde and Vali (2001a) for high-density plate crystals in the size range 0.2–1.5 mm or by Williams et al. (2011) are relatively uncommon in our geographical region for the type of meteorological events considered. Actually, Wolde and Vali (2001a) observations of dendritic crystals with sizes up to 4–6 mm ($\sim$2 dB at $-13^\circ\text{C}$) refer to nimbostratus (Ns) observations, while the highest observed $Z_{\text{dr}}$ (up to 7 dB) refer to hexagonal plates in shallow altocumulus (Ac), which are typically not associated with surface precipitation. We note in this context that the 5-mm threshold on the surface rainfall adopted as a selection criterion for this study led us to select a number of events characterized by similar warm-frontal synoptic conditions, especially relevant for their hydrological impact. While the mesoscale forcing associated with a warm front passage is expected to produce large areas of weak updrafts, the vertical velocity field is certainly modulated by smaller-scale turbulence, as evidenced by the frequent insurgence of the $Z_{\text{dr}}$ patchy patterns shown in Fig. 7. Outside the weak updraft regions it is then reasonable to expect that the air saturation over water is not always reached, making the environment favorable to hexagonal plate crystals (Magono and Lee 1966; Bailey and Hallett 2009). However, because of the much smaller sizes of these higher-density crystals, and considering the sensitivity limitations of our centimeter-wavelength radar (section 2), the fact that we did not observe in the cases analyzed a correspondingly high $Z_{\text{dr}}$ cannot be considered an indication of the lack of hexagonal plate crystals in these clouds.

Korolev et al. (2000) provided statistics from four measurement campaigns in stratiform clouds. Although the “plate” category was not included in that study (because of the insufficient resolution of the particle measuring system), the frequency of occurrence in the temperature interval from $-10^\circ\text{C}$ to $-15^\circ\text{C}$ of the largest particles (>500 μm, most likely to influence radar measurements) is shown to be dominated by irregular crystals (particles having an irregular or random shape—64%), followed by dendrites (28%). Given this general partition, it is not surprising to observe in our region the average $Z_{\text{dr}}$ distribution shown in Fig. 15.

b. $K_{\text{dp}}$ enhancement in the ice region

Figures 16 and 17 show two specific daily vertical profiles of reflectivity and $K_{\text{dp}}$ for stratiform (case study of section 3) and convective events, respectively. While in the convective case (Fig. 17) the $K_{\text{dp}}$ profile in the ice region is on average uniformly close to 0° km$^{-1}$, in the stratiform case (Fig. 16) the $K_{\text{dp}}$ enhancement around the $-15^\circ\text{C}$ level is well pronounced, especially when considering the 90th percentile. The $K_{\text{dp}}$ peak appears close to a local maximum in the height derivative of the reflectivity vertical profile. According to Lo and Passarelli (1982), the increased height derivative of reflectivity may be interpreted as an indication of the transition from vapor deposition to aggregation processes. Although aggregation normally occurs at temperatures warmer than $-5^\circ\text{C}$, a secondary maximum between $-10^\circ\text{C}$ and $-16^\circ\text{C}$ may exist when the arms of the dendritic crystals become entangled (Houze 1993). The decreasing skewness of the $K_{\text{dp}}$ distribution below the $-15^\circ\text{C}$ level (the median and the mean profiles get

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**Fig. 15.** Scatterplot of hourly averages $Z_{\text{dr}}$ vs $Z_{\text{dr}}$ at $-15^\circ\text{C}$ for stratiform precipitation at C band, with gray shades representing the associated $K_{\text{dp}}$ value. The vertical lines are drawn between the 10th and 90th quantile of the $Z_{\text{dr}}$ distribution.
closer toward the 0°C level, e.g., in Fig. 16) is also another repeatable feature observed in the vertical profiles, supporting the idea that the highly oblate crystals are being depleted by the aggregation taking place below the dendritic growth zone.

Figure 18 shows the temperature distribution of the maximum $K_{dp}$ location in the ice region ($T < 0°C$) for the profiles classified as stratiform. The 5°C-interval histograms show both the temperature distribution of the peak values derived from the daily average $K_{dp}$ (light bars) and from the daily 90th percentile (dark bars). The peak of the 90th percentile is well defined and centered at $-14.5°C$ (mean value), while the average $K_{dp}$ shows a less pronounced peak around $-12°C$. This is a clear indication that positive values of $K_{dp}$, when observed in the ice portion of the precipitation system, are preferentially associated with the dendritic growth zone around $-15°C$, as already pointed out in the case study discussion of section 3. The average $K_{dp}$ profiles are smoothed by averaging process and a local peak may not be easily recognizable. In fact, processes like growth by vapor deposition, as well as riming by...
small supercooled droplets when the air is saturated over water, are enhanced by updrafts, which normally occur over a confined space–time portion of the stratiform precipitation event. The secondary maximum in the temperature range from −5°C to 0°C of the histogram (average $K_{dp}$, light bars) arises from weaker precipitation events (the corresponding low-level mean reflectivity is 27.6 dBZ, as opposed to the averaged $Z_h$ of 30.6 dBZ for the cases where the maximum $K_{dp}$ is at colder temperatures), with $K_{dp}$ not showing a local maximum, but decreasing monotonically above the freezing level.

c. Statistical correlations between observations at −15°C and lower levels

Using particle trajectory calculations, Kennedy and Rutledge (2011) have shown the linkage between the positive $K_{dp}$ aloft and the snowfall intensification at the surface. While their conclusions were limited to snow, from a hydrological point of view it is extremely valuable to identify specific radar observations that may represent a precursor for the surface rainfall onset or intensification. We therefore exploit the large dataset of radar vertical profiles elaborated for the present study, seeking for statistical correlation between the radar observations aloft (specifically the $K_{dp}$ enhancement) and near the surface.

For the following analysis only events in which the freezing level was above 1.5 km MSL are considered (dark gray bars in the histogram of Fig. 13) to avoid vertical gradients associated with melting. This is done to allow meaningful comparisons with the lowest-level (900 m MSL) radar averages of $Z_h$ and $K_{dp}$, which are taken as representative of the mean rainfall rate near the surface. Among the initial 54 rainy days, only 4 were discarded based on the above threshold for the freezing level. We denote as $r(P^{cold}, Q^{warm})$ the Pearson correlation coefficient between a radar measurement $P$ at the upper cold level of −15°C ($P^{cold}$) and the measurement $Q$ at warmer temperatures ($Q^{warm}$, where warm denotes temperatures above −15°C at lower heights). Specifically, we are interested in analyzing $(P^{cold}, Q^{warm})$ pairs such as $(Z_h^{cold}, Z_h^{warm})$, $(K_{dp}^{cold}, K_{dp}^{warm})$, $(K_{dp}^{cold}, Z_h^{warm})$, where $K_{dp} = 10 \log_{10}(K_{dp})$ is used to linearize the relation between $Z_h$ and $K_{dp}$ (section 3b). Note that this procedure is not directly affected by the change in the dielectric factor among profiles since the correlation is calculated between two radar variables, each at a given temperature level, for example, $K_{dp}$ at −15°C (cold) and $Z_h$ at +5°C (warm). However, some minor influence on the correlation at different levels is possible since the change in the dielectric factor affects the radar detectability of particles.

In general the microphysical processes relevant for the precipitation growth depend on the temperature, and not on the absolute height. Therefore, we want to compute statistics as a function of the air temperature (taken from the COSMO local area model) to compare vertical profiles from different meteorological events. To compensate for varying surface temperature, the temperature at altitudes below the melting level ($T > 0°C$) is normalized to the average (over all events) lowest-level temperature of 12°C, according to

$$
\bar{T} = 12 \frac{T}{T_{900m}°C}.
$$

Figure 19a shows a scatterplot of the −15°C hourly reflectivity ($Z_h^{cold}$) versus the lowest-level reflectivity ($Z_h^{warm} = Z_h^{9°C}$) for convective (open circles) and stratiform (filled circles) events. The overplotted regression lines with confidence intervals emphasize the significant positive correlation between the reflectivity aloft and near the surface for both stratiform and convective precipitation. Figure 19b summarizes the computed correlation values for all the temperature levels and the $r(Z_h^{cold}, Z_h^{warm})$ correlation, with 3°C spacing, for both hourly and daily profiles. In Fig. 20 the analogous $r(K_{dp}^{cold}, K_{dp}^{warm})$ correlation is presented. The correlation is calculated on a total of 526 hourly profiles (260
and 266 convective) and on the 50 daily profiles. Although the daily profiles form a much lower statistical sample (23 stratiform and 27 convective events), the same qualitative results are found. The statistical significance of the correlations between the $-15^\circ C$ level and the lower levels subsequently reported is always ensured at the 0.5% level ($\alpha = 0.005$ is the significance level under which the null hypothesis of no correlation is rejected) for both hourly and daily stratiform profiles.

Reflectivity shows a decreasing correlation from the initial value of unity at $-15^\circ C$ (correlation between the $-15^\circ C$ level reflectivity with itself) to approximately 0.4–0.6 at the lowest level. There is a qualitative difference between stratiform and convective cases, with lower correlation in the melting layer for the stratiform precipitation and a more linear trend for convection. This behavior, and the higher slope of the regression between $Z_h$ aloft and near the surface (Fig. 19a), are explainable in terms of the lower stratification (higher mixing) for the convective cases and keeping in mind that the small-scale variability if filtered out by space–time integration in the calculation of the mean profile. No significant differences are observed between the 1- and 24-h correlation profiles. This result is indeed expected, since it forms the basis for the widely used vertical profile of reflectivity (VPR) correction methods (Koistinen 1991; Vignal et al. 1999).

Less predictable is the $r(K_{dp}^{\text{cold}}, K_{dp}^{\text{warm}})$ correlation (Fig. 20). For convective cases the correlation decreases rapidly to approximately zero below the melting layer, indicating that the specific differential phase in the rain medium has, in general, no relation with the values

![FIG. 19. (a) Scatterplot of hourly average $Z_h$ aloft ($-15^\circ C$) vs $Z_h$ at the lowest level (900 m MSL) for stratiform and convective precipitation. The solid black (gray) line is the regression for stratiform (convective) cases; the dashed lines mark the 99% confidence interval. (b) Correlation coefficient between $Z_h$ aloft ($-15^\circ C$) and $Z_h$ at lower levels ($3^\circ C$ spacing) for both hourly and daily average profiles. The asterisk superscripts on the $y$ axis temperatures indicate normalized values [Eq. (1)].](image)

![FIG. 20. Correlation coefficient between log-transformed $K_{dp}^{\text{aloft}}$ and $K_{dp}^{\text{lower levels}}$ for both hourly and daily average profiles.](image)
observed aloft in the ice region. On the other hand, for stratiform cases, the $r(K_{dp}^{cold}, K_{dp}^{warm})$ correlation shows a behavior closer to the $r(Z_{h}^{cold}, Z_{h}^{warm})$ correlation, with moderate values (>0.5) at all temperature levels.

For stratiform events the measured range of the hourly 90th-percentile $K_{dp}$ aloft over all the events is 0.02 to 0.99 km$^{-1}$, with 8% of the values of $K_{dp} < 0.1$ km$^{-1}$. On the other hand, the range of hourly 90th-percentile $K_{dp}$ at the lowest 900-m height level is 0.04–0.59 km$^{-1}$, with a higher fraction (31%) below the 0.1 km$^{-1}$ threshold.

We notice that the correlation of $K_{dp}$ aloft with the corresponding $K_{dp}$ at the lower levels could be affected by the lack of sensitivity of $K_{dp}$ measurements in the light rain, such as that typically occurring in light rainfall intensities of stratiform precipitation systems. For this reason it is useful to consider also the $r(K_{dp}^{cold}, Z_{h}^{warm})$ correlation, that is, the correlation between $K_{dp}$ aloft at the −15°C temperature level and $Z_{h}$ at lower altitude levels, since $Z_{h}$ has no such sensitivity limitation within the 50-km range.

A scatterplot of the hourly average $K_{dp}$ at the −15°C temperature level versus $Z_{h}$ at the lowest 900-m level, with regression lines for stratiform and convective events, is shown in Fig. 21a. The regression line over the stratiform sample shows a positive slope indicating an increase in $K_{dp}$ with increasing $Z_{h}$. On the contrary, the points representing convective cases appear mostly randomly distributed with a weak negative correlation. Figure 21b shows the $r(K_{dp}^{cold}, Z_{h}^{warm})$ correlation summary for all levels. For stratiform cases the correlation varies between approximately 0.5 above the freezing level and 0.7 below, for both hourly and daily profiles.

If we enforce a stricter definition of stratiform precipitation, considering only events with RCP < 25th percentile, then the correlation between $K_{dp}$ at −15°C and $Z_{h}$ at 900 m increases to approximately 0.8 for both hourly and daily profiles, while it remains around 0.5–0.6 between $Z_{h}$ at −15°C and $Z_{h}$ at 900 m (not shown). This marked correlation can be seen as an indication of the ability of $K_{dp}$ aloft to represent the ice water content (IWC) [Bringi and Chandrasekar 2001, their Eq. (7.17)]. Since we are applying a relevant space–time averaging in the derivation of the vertical profiles, $Z_{h}$ in the rain layer is well related to the precipitation liquid water content (LWC). In fact a Marshall–Palmer exponential relation represents fairly well the drop size distribution (DSD) after enough space–time averaging (Yangang 1993). For ice PSD mainly arising from water vapor deposition (platelike crystals) the IWC has been shown to be linearly related to $K_{dp}$, as opposed to the power-law relation between the IWC and $Z_{h}$ (Vivekanandan et al. 1994). In the absence of other microphysical processes contributing to IWC generation at lower altitudes (riming, accretion) and neglecting evaporation, the LWC below the freezing level is expected to be strictly connected to the IWC around −15°C. This is because the particle distribution undergoes modifications mostly through aggregation in the layer between −15°C and 0°C and therefore should keep its IWC essentially unchanged. The correlation between $K_{dp}$ aloft and $Z_{h}$ in
rain is then taken as an evidence of a plausible representation, on average, of the IWC by \( K_{dp} \) in this particular ice region of the stratiform clouds.

The results shown in Figs. 16–21 clearly reveal the existence of a statistically significant link (at the 0.5% significance level) between peak values of \( K_{dp} \) in the ice region and the precipitation near the surface for mesoscale widespread precipitation systems. No positive correlation is found for convective precipitation, for which other processes such as heavy riming may overwhelm the depositional growth of ice. A slightly negative correlation is found actually for convective events (Fig. 21b), more pronounced \((r' \sim -0.3)\), but less significant \((15\% \text{ significance level})\) for daily average profiles. This negative correlation, opposite to the stratiform case, can be attributed to the increasing role of particle growth by riming, which contributes to generate higher-density particles with lower \( K_{dp} \) in the ice region and larger \( Z_h \) over the whole vertical profile, resulting in higher rainfall intensity near the surface.

Unlike \( K_{dp} \), the reflectivity aloft was shown to be fairly correlated \((r \sim 0.5)\) with the lowest-level reflectivity for both stratiform and convective precipitation (Fig. 19). The \( K_{dp} \) peak around the \(-15^\circ C\) temperature level appears to be therefore a distinguished feature of stratiform precipitation, where the vapor deposition mechanism, possibly combined with condensational growth of liquid droplets in the water saturated cloud, can be considered the dominant ice-particle growth processes (Rogers and Yau 1988; Korolev 2007).

5. Discussion and conclusions

Differential phase shift measurements and specific differential phase estimators have received considerable attention in recent years mainly because of their potential for hydrological applications (Ryzhkov and Zrnić 1996; Gorgucci et al. 2000). The \( K_{dp} \)-based rainfall rate estimators in particular have become increasingly popular, especially at attenuating frequencies such as C band and X band, for their insensitivity to path attenuation and the increased sensitivity for light rain measurements at these frequencies (Wang and Chandrasekar 2010). In contrast, measurements in the ice region of precipitating systems where the hydrometeors can exhibit considerable anisotropy to dual-polarization radar methods have been only occasionally considered (Hendry et al. 1976; Vivekanandan et al. 1994; Ryzhkov and Zrnić 1998; Trapp et al. 2001; Kennedy and Rutledge 2011). The present study has extended the results of Kennedy and Rutledge (2011) who analyzed \( K_{dp} \) at S band in several winter storms in Colorado. For the present research a large dataset of measurements collected between 2009 and 2011 by the operational C-band radar of Bric della Croce (Turin, Italy) was processed. Additionally, a single significant stratiform event observed by the C-band radar and by the nearby deployed ARX transportable X-band system has been studied in more detail.

A first result of the present work elucidates the microphysics behind the studied \( K_{dp} \) signatures in stratiform precipitation. The analysis of the coincident C-band and X-band measurements in the ice region demonstrated that, in agreement with electromagnetic scattering simulations, the ice particles in the region centered around \(-15^\circ C\) are Rayleigh scatterers because \( K_{dp} \) scales with frequency. Vapor deposition, associated with riming by small supercooled droplets at water saturation (Korolev 2007), is the most relevant snow growth process in stratiform clouds, occurring in the region where the difference between the saturation vapor pressure over water and the saturation vapor pressure over ice is the greatest. In this region, between approximately \(-12^\circ\) and \(-16^\circ C\), platelike crystals present the most effective shape for the deposition of the ambient water vapor by virtue of the high surface-to-volume ratio (Houze 1993). The ice crystal habit is in general a function of both temperature and supersaturation (with respect to ice). In particular, the appearance of hexagonal plate crystals (high density) or dendrites (lower density) in the range from \(-10^\circ\) to \(-20^\circ C\) is governed by the relative humidity and the consequent level of supersaturation in the cloud (Bailey and Hallett 2009). As noted in Williams et al. (2011), dendritic crystals represent the most likely crystal habit for the considered temperature range in water saturated regions. A detailed analysis of the polarimetric radar signatures for a single case showed qualitative and quantitative agreement with previous observations of dendritic crystals (Wolde and Vali 2001a). The following evaluation of the polarimetric characteristics of the average vertical profiles for 27 days of stratiform precipitation allowed to extend this evidence and to deduce that dendrite crystals represent a common crystal habit in the \(-15^\circ C\) region, for autumn and spring stratiform precipitation in the Italian subalpine region.

The presence of dendritic crystals, whose branches are prone to become entangled (Houze 1993) and collect the available supercooled droplets to form rimed aggregates, is probably one plausible reason for which these signatures are so easily revealed by centimeter-wavelength radars (Andric et al. 2009; Kennedy and Rutledge 2011; Williams et al. 2011), whose detection capabilities are strongly dependent on the availability of large scatterers. This leads us to remark that, although dual-polarized radar observations provide an invaluable
tool for the observation of ice clouds, the crystal habits and relevant microphysical processes can only be comprehensively depicted using a wider range of observing instrumentation systems, ultimately relying on in situ measurements.

Second, statistical evidence is presented that the $K_{dp}$ signature in stratiform precipitation occurs near the $-15^\circ$C level, irrespective of altitude. Being independent of the precipitation type at the ground, it appears to be a relevant feature not only in snowfall, but more generally for widespread rainfall events, which in the area of interest are most likely to occur during spring and fall. For the 27 days analyzed the height of the $-15^\circ$C temperature level ranged between 3.9 and 6.0 km MSL and $Z_h$ at the lowest level for 10 equal-sized classes of the RCP. The values on the x axis at the base of the bars denote the left and right limits of the RCP classes.

The skewness of the reflectivity distribution is considered to define a convective parameter named RCP (described in the appendix). More generally, the skewness of the distributions of the polarimetric variables is an important characteristic that could be further exploited to derive valuable information about the microphysical state and processes of precipitation particles. A local characterization of the distributions, for example, by $n$-quantiles calculation over limited space–time domains, may also have a potential to benefit the performance of hydrometeor classification schemes (Vivekanandan et al. 1999; Liu and Chandrasekar 2000).

Figure 22 finally summarizes the correlation statistics ($K_{dp}$ aloft versus the near-surface reflectivity) as a function of the RCP value associated with the analyzed hourly profile. In this plot the Pearson correlation $r$ and the rank (Spearman) correlation $r_S$ are shown for 10 equal-sized classes, each bar representing the correlation calculated over approximately 50 samples. The rank correlation does not assume a linear relationship, but just a monotonic one, so there is no need to linearize the $K_{dp}$–$Z_h$ relation by taking the log values of the former variable, as for the Pearson correlation. The rank correlation is in general higher than the Pearson correlation, as expected, being more robust to outliers (Myers and Well 2003). The plot highlights the dramatic difference between stratiform and convective precipitation. In particular, the trend presents a marked step change around RCP = 4 dB, with the lowest 40% of the total sample having a rank correlation near $r_S = 0.8$. This presents clear evidence that the statistical correlation between the $K_{dp}$ signatures associated with snow growth aloft and the surface precipitation rate is unique to stratiform precipitation.

The correlation statistics focused on events with a relatively high freezing level (>1.5 km MSL) for ease of comparison with the low-level reflectivity, representative of the surface precipitation rate. While precipitation associated with purely stratiform rain has generally moderate consequences on the subalpine Italian territory, even weak snowfall events over the plains can have a major impact, especially on ground transportation and aviation. For this reason it will be fundamental to assess whether and under which circumstances the presented results may extend to snowfall episodes. At the same time, the origin of the statistical correlation between observations aloft and surface precipitation will need to be further explored. In particular time-lagged analysis between $K_{dp}$ in the ice region and surface precipitation (from either gauges or radar) will help to assess if there is a potential for prognostic applications.
Appendix

Radar Convective Parameter

Following Houze (1993), stratiform precipitation is defined in terms of the vertical air motion leading to the formation of precipitation particles, that is:

\[ w \ll V_{\text{ice}}, \quad (A1) \]

where \( w \) is the vertical air velocity and \( V_{\text{ice}} \) is the terminal fall speed of ice precipitation particles (\( \sim 1-3 \text{ m s}^{-1} \)). Conversely, the precipitation type is defined to be convective when (A1) does not hold. Unfortunately, observations of the vertical air velocity are scarcely available, so a definition in terms of the radar reflectivity is needed for many applications. The condition (A1) implies that all hydrometeors are constantly falling, therefore avoiding larger and denser particles growing by aggregation and riming to be found in the upper atmospheric levels. On the other hand, in convective precipitation, the relationship between upward vertical air velocity and the development of ice condensate aloft results in much larger particles at subfreezing temperatures (Williams et al. 1992). The main consequence of the limited vertical air velocity in stratiform precipitation is the highly stratified reflectivity structure above the melting layer. The subsequent melting of the ice crystals and aggregates leads to the characteristic “brightband” signature, whose presence (absence) is often used to classify a reflectivity profile as stratiform (convective). The brightband-based classification is dichotomic; that is, it does not allow a continuous parameterization of the degree of convectivity. In addition the detection of the bright band is typically based on several empirical thresholds on the reflectivity values and gradients, which are also sensitive to the radar operation frequency because of complex electromagnetic scattering properties of mixed-phase particles. For the scope of this study we wanted instead to perform an objective continuous classification of the average reflectivity profiles, either daily or hourly. We therefore defined the RCP, a simple parameter to describe the degree of convection in a given reflectivity vertical profile, as

\[ \text{RCP(dB)} = 10 \log_{10} \left[ \int_{h(0^\circ \text{C})}^{h(-15^\circ \text{C})} \frac{\langle Z_{\text{lin}} \rangle}{\text{median}(Z_{\text{lin}})} \, dh \right] \quad (A2) \]

where \( \langle Z_{\text{lin}} \rangle \) [the average of the reflectivity in linear units (\( \text{mm}^6 \text{ m}^{-3} \))] and median\( (Z_{\text{lin}}) \) are calculated over all observations available at a given vertical level \( h \) and the integral is over the atmospheric layer bounded by the \( 0^\circ \text{C} \) and \(-15^\circ \text{C} \) temperature levels. In practice the integral in (A2) is replaced by a summation over the available discrete height levels. As an example, the calculated RCP values for the stratiform and convective profiles in Figs. 16 and 17 are, respectively, 3.8 and 16.6 dB.

The summation is performed on the ice portion of the vertical profile only, despite the fact that higher reflectivity differences between convective and stratiform situations are expected below the \( 0^\circ \text{C} \) level (Steiner et al. 1995). Likewise the reflectivity enhancement in the bright band is not involved in the definition of the RCP to eliminate the variable depth of the liquid and mixed-phase layer, especially when dealing with a large dataset encompassing a wide range of freezing-level heights (0.4 to 4.3 km MSL for the profiles analyzed in this study).

It should be noticed that (A2) is actually very similar to the definition of the Pearson’s second skewness coefficient:

\[ \text{skewness} = 3 \frac{\text{mean} - \text{median}}{\text{standard deviation}}, \quad (A3) \]

The main difference is that in (A2) there is no normalization by the standard deviation. The higher vertical air motions in convective precipitation allow mixed-phase particles (graupel) and supercooled liquid water (with approximately fivefold higher refractive index relative to ice, for the 5–10-GHz frequencies considered in this work) to reach heights with temperatures below freezing. The presence of scatterers with higher equivalent reflectivity values in convective situations contributes to broaden the distribution, whose variance increases. The difference between the mean and the median also notably increases because of the highly positively skewed distribution of the reflectivity in linear units. The use of (A2), relative to (A3), is therefore preferred to enhance the dynamic range of the convective parameter.

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