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# THE USE OF CO-POLAR CORRELATION COEFFICIENT TO SEPARATE REGIONS OF RAIN, MELTING LAYER AND SNOW REGIONS FOR APPLYING CORRECTIONS TO VERTICAL PROFILES OF REFLECTIVITY

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## 1. INTRODUCTION

Operating weather radars in a mountainous terrain typically requires the use of higher radar elevation angles because of the partial or complete beam blockage at low elevations. Coupled with the low heights of the freezing level, this observational geometry causes that radar resolution volumes are often located in the melting layer or snow regions even if it rains at the ground.

One of the approaches proposed to account for this when applying radar reflectivity - based rainfall rate estimators is to use a priori information about the vertical profiles of reflectrivity (VPR; e.g., Andrieu and Creutin 1995; Bellon et al. 2005; Koistinen, 1991; Kitchen 1997). According to the VPR approach, the reflectivity measured aloft in the snow or melting hydrometeor regions is related in a mean sense to an expected reflectivity at the ground in the rain region.

This presentation describes a modified VPR approach for quantitative precipitation estimations (QPE). This approach is specifically tailored to the NOAA ESRL X-band radar polarimetric radar measurements, however, the general concept also applies to polarimetric radars operating at other wavelengths. In traditional VPR approaches a priori information (e.g., climatological values) about the heights of the melting layer is usually used. The main feature of this approach is that the polarimetric measurements are used to identify the location and the extent of the melting layer for each particular slant (i.e., low elevation angle) radar beam, thus a significant source of errors associated with uncertainties of locating the melting layer boundaries is eliminated. Eliminated are also uncertainties associated with beam broadening effects, which results in an apparent increase of the melting layer thickness as the range increases.

Because the polarimetric QPE in mixed phase hydrometeors (i.e., in the melting layer) and snow remains largely unexplored area, X-band radar polarimetric capabilities are used here mainly for determining melting layer boundaries and to correct reflectivities for attenuation at shorter ranges that are filled with rain. Mean reflectivity – based estimators are used for QPE once reflectivity values at the ground are estimated using mean VPR and beam-specific information on locations of melting layers.

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## 2. OBSERVED AND MODELED X-BAND VPRs

Figure 1 shows examples of vertical profiles of reflectivity measured with a vertical radar beam (a) and reconstructed from RHI scan measurements at an 18 km distance from the radar location (b). These X-band radar measurements were conducted in a steady rain ($R \sim 3$ mm h$^{-1}$) observed on 2 January 2005 in the field experiment in the California Sierra Nevada foothills. Figure 1a also shows the model simulations of the bright band (BB) reflectivity enhancement in the melting layer calculated using the Wiener (1910) mixing rule for dielectric constants of melting particles.

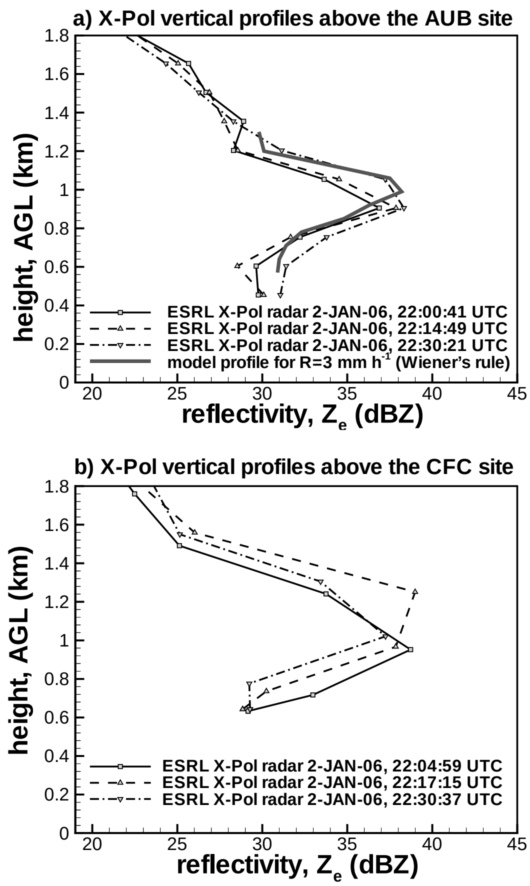


FIG.1. Vertical profiles of reflectivity for vertical beam measurements (a) and from RHI data at an 18 km distance during a steady rain

Vertical profiles of reflectivity reconstructed from RHI scans at different distance from the radar location were used to construct a mean idealized VPR. This mean profile, that was adopted further in this study for quantitative precipitation measurements, is shown in Fig. 2.

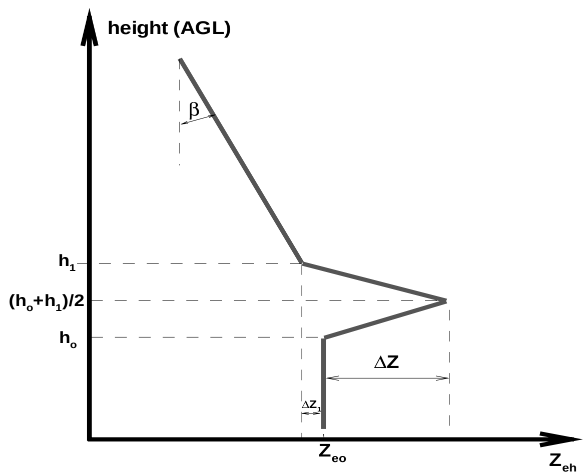


FIG.2. An idealized mean vertical profile of reflectivity in a stratiform rain

In Fig.2, $\Delta Z$, $Z_{eo}$, tan (β), $h_o$, and $h_1$ are the reflectivity BB enhancement, the mean reflectivity in the rain layer, the mean reflectivity gradient in the in the snow region, and the bottom and the top heights of the melting layer, respectively. Note that $h_1$ corresponds also to the altitude of the freezing level.

Model calculations show that, in the idealized case, the BB reflectivity enhancement does exhibit strong dependence on rain rate below the melting layer. However, beam broadening effects and also attenuation in rain result in gradual diminishing of $\Delta Z$ as the distance from the radar increases. When the beam broadening and the attenuation effects are small, $\Delta Z$ is about 6 -7 dB for typical stratiform precipitation events when riming of snowflakes above the freezing level is small. The value of $\Delta Z_1$ representing the difference between reflectivities in the rain layer and in the snow region just above the freezing level generally changes as a function the attenuation of radar signals in the melting layer. For radar elevation angles greater than 3$^o$ and rain rates less than about 5 mm h$^{-1}$ , $\Delta Z_1$ is about 2 dB and it varies relatively little. A typical value of the reflectivity gradient in the snow region above the freezing level [i.e., tan (β)] is about 5 dB km$^{-1}$.

## 3. IDENTIFYING MELTING LAYER BOUNDARIES

In order to apply a VPR correction to reflectivity measurements, a position of the radar resolution volume with respect to the melting layer boundaries needs to be established. The melting layer typically manifests itself by the reflectivity increase. However, positioning the BB along a slanted low-elevation radar beam is challenging because horizontal variability of rainfall (and thus the horizontal variability in reflectivity) is often significant. This variability can considerably complicate any automatic procedure of determining melting layer boundaries.

The co-polar correlation coefficient $\rho_{hv}$ is a very useful parameter that allows a relatively robust discrimination among the regions of rain, melting and snow. The magnitude of this coefficient is related to the variety of hydrometeor shapes present in the radar resolution volume. For an idealized situation of the ensemble of identical particles, $\rho_{hv}$ is unity. The experimental value of $\rho_{hv}$ observed with the NOAA ESRL X-band radar in rain is generally greater than 0.95 -0.96 and it does not practically depend on rain intensity. In snow, $\rho_{hv}$ values are generally greater than 0.85-0.9. Values of $\rho_{hv}$ observed in the melting layer are significantly smaller than those that are typical for the rain and snow regions. This fact allows a relatively straightforward way of identifying melting layer boundaries (along a slanted radar beam) as the areas where measured $\rho_{hv}$ undergoes transitions between high and relatively low values.

Figure 3 shows two examples of the low elevation (3$^o$) slant beam radar measurements. As nicely indicated by the $\rho_{hv}$ transitions from high to low (and vice versa) values, the melting layer is confined between slant ranges of about 23 and 33 km (Fig. 3a), and 16 and 24 km (Fig. 3b).

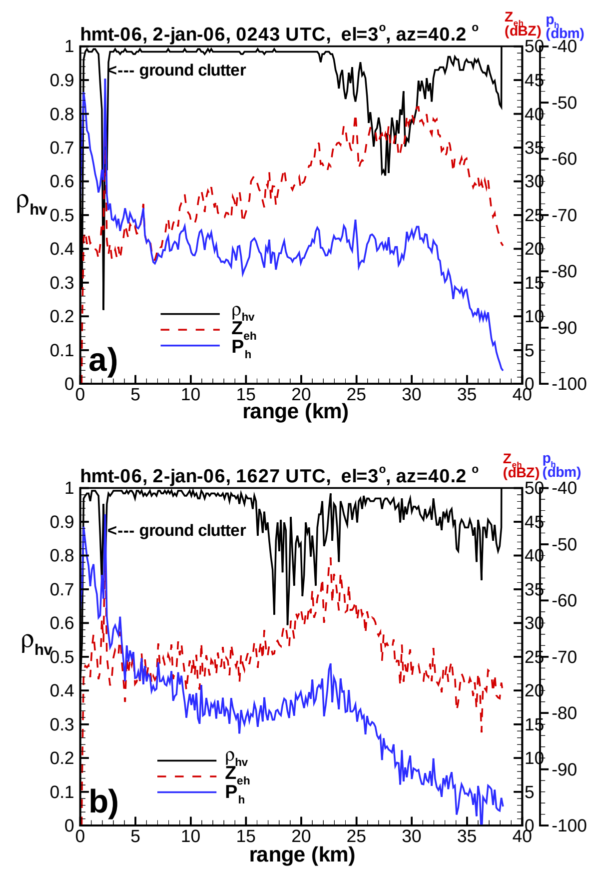


FIG.3. Slant-beam range dependence of co-polar correlation coefficient $\rho_{hv,}$ reflectivity corrected for attenuation in rain $Z_{eh}$, and the power on antenna terminals $P_h$.

## 4. APPLYING THE VPR APPROACH

An automatic procedure based on $\rho_{hv}$ measurements was developed to determine BB boundaries. As son as the slant ranges of the melting layer (i.e., the reflectivity BB) are established for each radar beam, The mean VPR shown in Fig. 2 is applied to all the ranges where the radar resolution volume is either in the melting layer or in the snow region. According to this mean VPR, measured reflectivities (that are routinely corrected for attenuation in the rain region using differential phase measurements) are re-calculated to the values that are expected to be in rain near the ground. It is assumed that the actual rain reflectivity does not change with height. This assumption is generally valid for stratiform rains. An example of applying the VPR method with the $\rho_{hv}$-based determination of the melting layer boundaries is shown in Fig.4.For these data, the elevation angle is 3°.

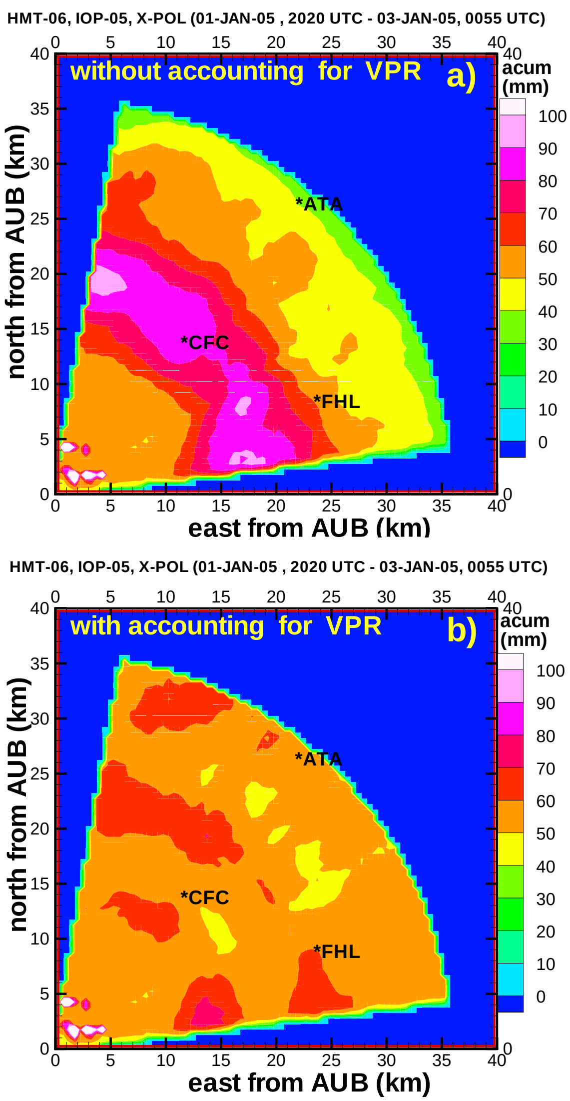


FIG.4. The total rainfall accumulation maps for the event 1-3 January 2006 without (a) and with the VPR correction.

Figure 4a shows the total event rainfall accumulation map obtained when no-VPR correction was introduced. An obvious manifestation of the BB is seen as an arc of increased accumulation values centered at a radius of about 19-20 km. Beyond the BB arc, the accumulation diminishes with range as the radar observes the snow region where reflectivity gradually diminish with height above the ground and, hence, with the range from the radar.

The accumulation map obtained with the VP correction applied is shown in Fig. 4b. Although there are still some features that can be regarded as the manifestation of the BB, they are much less pronounced relative to those in Fig. 4a. The map obtained with the VPR correction is more homogeneous, and it does not have area with vastly different accumulation values. Accumulations in Fig. 4b are in much better agreement with the gauge data than those in Fig. 4a.

## 5. CONCLUSIONS

The VPR correction procedure significantly improves high-resolution quantitative precipitation estimates. Especially important are these improvements when radar operations are performed under conditions of low freezing levels which are common for recent operations with the NOAA ESRL X-band scanning polarimetric radar (Matrosov et al. 2005). Unlike in many other VPR correction schemes, the exact location of the reflectivity bright band (i.e., the boundaries of the melting layer) is estimated on a beam-to-beam basis using measurements of the co-polar correlation coefficient between horizontally and vertically polarized radar signals. This coefficient provides a robust discrimination among regions of rain, melting hydrometeors, and snow along the beam. An important advantage of this correlation coefficient is also that $\rho_{hv}$ threshold levels corresponding to phase transitions practically do not depend on precipitation intensity.

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