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Vertical profiles of radar reflectivity and rain rate: observational and modeling results

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Abstract. Implications for quantitative precipitation estimation coming from characteristics of mean radar reflectivity and rain rate profiles are investigated for the case of a two–day stratiform rain event in central Oklahoma. Recent analytical findings on height–dependent correction of measured reflectivity to eliminate effects of vertical air density stratification on rainfall estimates are evaluated. The main findings are that at large range of more than about 100 km precipitation accumulation can be improved significantly with the proposed procedure. However, all artifacts on the measured reflectivity field have to be removed before Z–correction can successfully be applied: In the radar data used, a clear melting layer signature was present in the data, and the exaggerated reflectivity in this region was further enhanced by our correction procedure. Therefore, the idea to operationally run bright band removal algorithms in quantitative precipitation estimation algorithms is supported by our analysis.

1 Introduction

When radars are applied for Quantitative Precipitation Estimation (QPE), the accuracy of the procedure is range–dependent, even if the radar is located on flat terrain, where no beam blockage or other orographic effects occur. However, in regions where a radar network is sparse, QPE may have to be extended to large ranges up to 250 km, or under special circumstances up to 300 km. Even a 0.5° base–level radar scan is several kilometers above the ground level (AGL) in this case. Due to the large beam height, correlation of radar reflectivity Z aloft and rain rate R at the ground is a major issue: First, due to lower air density aloft, R and Z decorrelate with height z, because R increases with height, while Z is a constant for a given hyrometeor distribution. Second, there is evidence that the vertical profile of reflectivity itself also decorrelates from its ground value even at a few kilometers AGL.

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As shown by Dotzek and Beheng (2001), the effect of lower air density aloft on R can be eliminated almost completely by a simple, air density–dependent correction factor on radar–measured reflectivity. After a mesoscale model study on the effects of convective vertical drafts on R and Z–R relations, now this effect of density stratification and its simple correction is being studied with real radar data. The goal of this investigation is to test (i) if this correction could significantly improve QPE accuracy at large radar ranges of more than 100 km, (ii) what is necessary to successfully apply the correction, and (iii) if further and more complex studies would be worthwhile.

The paper is organized as follows: Sect. 2 reviews the different data used in this study. Sect. 3 describes an air density–dependent correction procedure to measured radar reflectivity and applies this to a 48 hour stratiform rain event in Oklahoma. Sects. 4 and 5 present discussion and conclusions.

2 Available data

2.1 QPE SUMS

Researchers at the National Severe Storms Laboratory (NSSL) in Norman, Oklahoma have developed a real–time QPE algorithm that utilizes data from the operational US Weather Surveillance Radars – 1988 Doppler (WSR–88D). The radar network employs approximately 120 radars that are jointly operated by the Department of Defense, Federal Aviation Administration, and National Weather Service. The aforementioned algorithm, formally called Quantitative Precipitation Estimation and Segregation Using Multiple Sensors (QPE SUMS, Gourley et al., 2001), has been designed in a flexible manner so that scientists can test new approaches to QPE. In its current, operational design, QPE SUMS ingests radar data from multiple WSR–88Ds, infrared satellite data from the Geostationary Operational Environmental Satellites (GOES), numerical model output, surface and upper air observations, and lightning flash data. After the radar

Table 1. System specifications of S–band precipitation profiling radar

Wavelength	10.6 cm, S–Band
Amplifier	solid state–low voltage, Class C
Peak Power	380 W
Max. Duty Cycle	10%
Antenna	3 m diameter shrouded parabolic dish
Beamwidth	3.2°
Height Resolution	selectable: 60, 100, 250, 500 m
Max. Height Sampled	21.4 km (typical)
Max. Radial Velocity	20 m s^{-1} (typical)
Spectral Points	128 (typical)
Dwell Time	35 s (typical)
Processor	DOS PC–based
Sensitivity	–25 dBZ$_e$ at 10 km altitude
Power Requirements	[110 or 220 $\pm$ 10%] V AC, 5 A
Recording	full Doppler spectra
Media	optical disk

data are quality–controlled for anomalous propagation and ground clutter contamination, precipitation echo is interrogated to determine if it is either stratiform or convective. This segregation determines which reflectivity to rainfall equation (Z–R) will be applied. The additional data sources (e.g., GOES satellite–derived brightness temperatures) are used primarily in stratiform rain situations where radar data are known to have problems due to bright band contamination and underestimation due to beams overshooting precipitation structures at intermediate and far range. Details of a prototype multisensor approach were presented by Gourley et al. (2002).

2.2 S–band profiling radar

During the winter seasons of 2000 and 2001, a profiling radar was employed in the mountainous terrain of northern Arizona, U.S.A. The operating characteristics of the radar are provided in Table 1. The vertically–pointing radar was sited in a prominent valley that receives most of its annual precipitation in the form of stratiform rain. One goal of this experiment was to determine in a statistical manner how reflectivity measured aloft relates to surface precipitation rates in stratiform rain. A wealth of research has shown the range dependence of radar precipitation estimates primarily due to the significant change in reflectivity with height (e.g. Kitchen and Jackson, 1993). In light of this, correction methods have been devised to reduce range biases by using an observed Vertical Profile of Reflectivity (VPR, Joss and Lee, 1995; Gysi et al., 1997). While VPR corrections will inevitably lead to improvements over uncorrected (range–biased) rain rates, questions still remain as to the correlation between reflectivity values measured at large heights to surface rain rates.

To address such questions, 3377 profiles of stratiform rain with a bright band observation were chosen for statistical analysis. A linear correlation coefficient is computed at each 43 m range gate up to 5 km ASL in the vertical. The correlation coefficient quantifies the relationship between reflectivity measured at each height bin above the radar to reflectivity measured at the lowest height bin of 1217 m ASL (268 m above radar elevation), which can serve as a proxy for surface rainfall.

Statistical results from this observational study are shown in Fig. 1a. The correlation coefficient r^2 (not to be confused with radar range r) decreases rapidly up to a height of approximately 3 km ASL. Above that height, the correlation coefficient asymptotically approaches zero, reaching its minimum value of 0.01 at the maximum observation height of 5 km ASL. Most notably, reflectivity measured aloft decorrelates with surface reflectivity at relatively low heights above the radar.

2.3 Model–derived average rain rate profiles

In a combined analytical and numerical study, Dotzek and Beheng (2001) as well as Dotzek and Fehr (2002) investigated the effects of deep convection on the applicability of standard Z–R relations. First, convective air motions in large cumulus clouds can be very strong, supporting even large hydrometeors at constant altitude or even lifting them in vigorous updrafts. Downdrafts, however, accelerate hydrometeors to the ground, thereby enhancing the precipitation rate. On the other hand, reflectivity Z is not influenced by vertical drafts directly. As a result, in large convective clouds, Z–R relations can take on almost any functional form, making application of standard Z–R relations questionable or even meaningless. As vertical drafts cannot be readily derived from single radar measurements, the reader is again referred to the work on Dotzek and Fehr (2002) which covers this problem with mesoscale model simulations.

Secondarily, R is also altered by reduced air density aloft, leading to larger terminal fallspeeds of hydrometeors. This effect can be nearly eliminated from the procedure to derive R at the ground from observed Z aloft. Sect. 3 will exemplify this in greater detail.

Neglecting vertical drafts, Fig. 1b from the MM5 model simulation of a multicell storm (Dotzek and Fehr, 2002) shows that horizontally averaged, storm–scale instantaneous vertical profiles of precipitation rate $\langle R\rangle(z)$ can show a similar decorrelation with height as did the average profile in Fig. 1a. Again, the largest decorrelation occurs within the lowest 3 km AGL. Even though the model simulation was made for thunderstorm clouds, while the S–band profiler radar observations were made in stratiform precipitation, the similarity is striking.

More extended investigations are necessary, however, to test the general validity of the vertical decorrelation in average $\langle Z\rangle$ and $\langle R\rangle$.

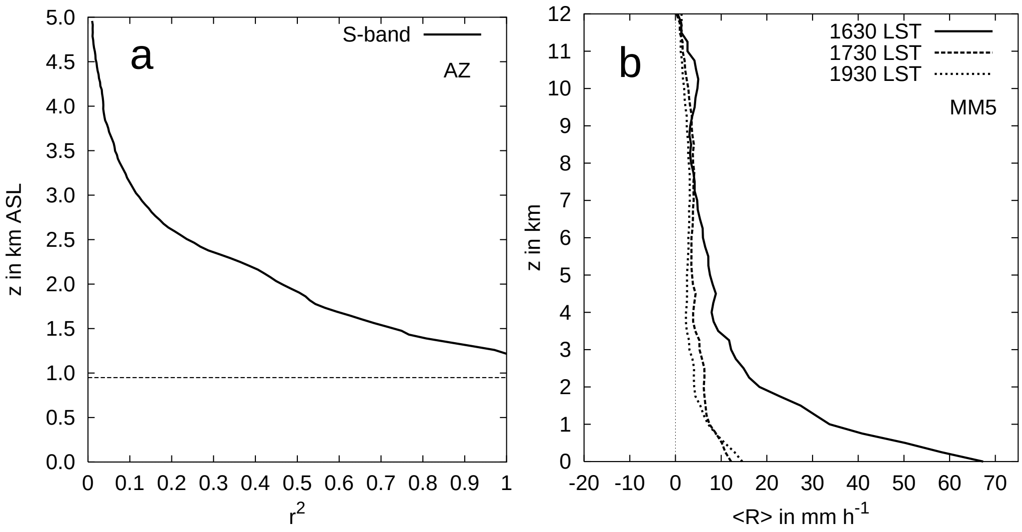

Fig. 1. Vertical correlation profile for Z at S–band profiling radar in Arizona (a), dashed line marks antenna heigth), and (b) horizontally averaged $\langle R \rangle$ profiles from an MM5 multicell storm simulation at three different times (Dotzek and Fehr, 2002).

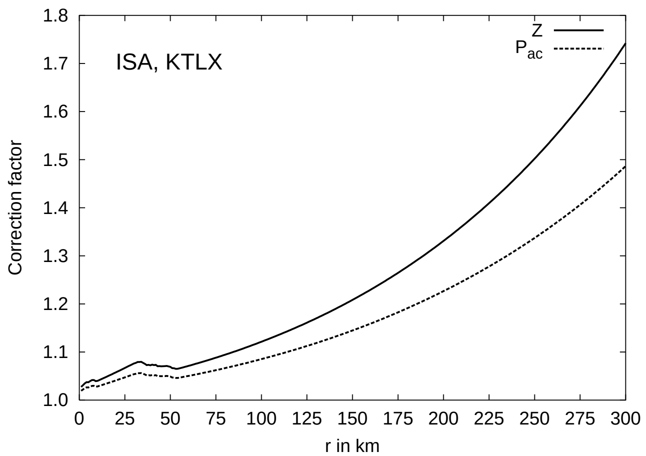

Fig. 2. Azimuth–averaged value for the Z correction factor (solid) and the resulting factor for precipitation accumulation P_{ac} at KTLX radar assuming ISA conditions.

3 Vertical reflectivity profile correction

3.1 Procedure

QPE SUMS is used here as a tool for ingesting WSR-88D radar data and performing quality control in order to test the effect of a height–dependent reflectivity correction factor. The module tested is designed to account for the increase in rain rates aloft due to reduced air density. Theoretical and modeling studies of this density effect on reflectivity–based rain rate computation have been investigated analytically in detail by Dotzek and Beheng (2001). QPE SUMS now allows us to evaluate the analytical results for real meteorological cases.

Neglecting ambient vertical air motions, and focusing on density stratification $\rho(z)$ from its sea level value ρ_{00} to smaller values aloft, the precipitation rate R reads in bulk variables

$$R = \bar{w}_{t,00} \left(\frac{\rho_{00}}{\rho} \right)^{\alpha} \rho\, q\,, \tag{1}$$

with $\rho\, q$ denoting hydrometeor mass per unit volume and $\bar{w}_{t,00}$ hydrometeor fall speed. The density stratification term was chosen in its most widely used form with an exponent $\alpha = 0.45$ (Beard, 1985).

The biggest question to answer is how important this density effect in Z–R relations can be when performing quantitative precipitation estimates. Locally, in convective clouds R is much more affected by strong vertical drafts. But when making radar observations at large heights in thunderstorms or with base–level scans at large (100 km) or very large (∼200 km) range, in order to compute hourly or daily precipitation accumulation, even this secondary effect becomes considerable.

As a physical density correction factor Dotzek and Beheng (2001) found for $Z = a\, R^b$ and the fall speed law exponent $\alpha = 0.45$ (Beard, 1985):

$$Z_{\rm corr} = Z \left(\frac{\rho_{00}}{\rho(z)} \right)^{\alpha\, b} . \tag{2}$$

Obviously, for precipitation aloft the measured radar reflectivity factor Z has to be increased according to this relation before any standard sea level Z–R relation can be applied to yield a representative rain rate R.

For the Twin Lakes WSR–88D radar (KTLX) only one standard Z–R relation was applied for this study, with coefficients $a = 300$, $b = 1.4$. To include the rain type segregation capabilities (and thereby apply local values for a and b) of QPE SUMS would be a matter for a future study. Figure 2 shows the azimuthal average of the correction factor

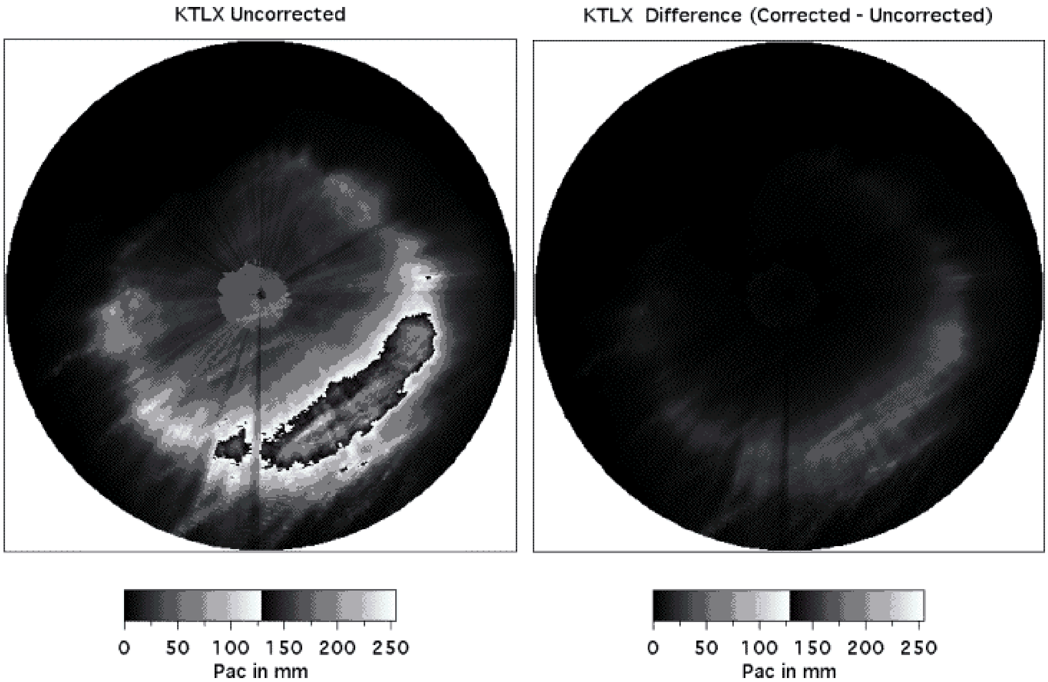

Fig. 3. Storm total precipitation from KTLX WSR–88D radar without Z–correction (left) and the difference field between ISA–corrected and uncorrected P_{ac}.

from Eq. (2) as a function of range for KTLX radar under the assumption of a density profile according to the International Standard Atmosphere (ISA, solid). Also shown is the resulting range–dependent enhancement of precipitation rate R at the ground and also precipitation accumulation P_{ac} (dashed). The steps in the two curves for radar ranges r below 50 km stem from changes in radar tilt angle. For $r > 50$ km only the base–level elevation angle of 0.5° is used for QPE.

The dense 120–rain gauge network from the Oklahoma mesonet is used to evaluate the effect of air density–based corrections. Data from the KTLX radar near Norman are ingested in the precipitation algorithm for a stratiform rainfall case in central Oklahoma. KTLX is a linearly polarized, 10 cm wavelength Doppler radar that measures reflectivity, radial velocity, and spectrum width at up to 14 elevation angles, along 360 radials (1° azimuthal resolution), out to a maximum range of 460 km (1 km range gate spacing).

3.2 Case study of stratiform rain

From 18 to 20 March 2002, a stratiform event of heavy precipitation occurred in central Oklahoma and lasted for roughly 48 hours. The center of the precipitation area was to the southeast of Norman, but close enough to apply data from the KTLX radar alone in this first study.

The radar data as well as the Oklahoma mesonet rain gauges were evaluated in the period from 12:00 UTC on 18 March to 12:00 UTC on 20 March 2002. Of the 120 rain gauges, 115 were operational in this period. While their data was available every 15 min, here we only look at the hourly precipitation accumulation. KTLX radar was operational the whole 48 h period, giving QPE scan products every 5 to 6 min up to a range of 300 km.

Weather conditions during this rain event were quite close to ISA conditions, and surface temperatures ranged between about 10 to 20° C. The KTLX data were quality–controlled for anomalous propagation and ground clutter. Any artificially enhanced reflectivity values due to the melting layer (“bright band”) were not removed, however. A melting layer signature was clearly present in the radar data of that period, the typical ring–like zone of high Z was obvious from the Plan Position Indicator (PPI) radar scans during the 48 h precipitation accumulation. The melting layer height varied with time over the 48 h period, therefore Z observations in the range interval from 150 to 200 km might be more or less contaminated by the melting layer signature. This range interval corresponds to roughly 2.5 to 4.0 km AGL for the 0.5° base–level PPI scans used for the QPE.

The region of maximum rainfall and the intersection of the KTLX radar beam with the melting layer nearly coincided southeast of the radar. Yet, as we are only interested in the relative increase of precipitation accumulation P_{ac} due to the reflectivity correction of Eq. (2), it is irrelevant if there is bright band contamination in the radar data. One should only keep in mind that a strong melting layer signature can lead to overestimation of rainfall by radar. In most other cases, at larger range, radars usually underestimate P_{ac}.

Figure 3 shows the uncorrected 48 h P_{ac} in the left panel, compared to the difference between ISA–corrected and un-

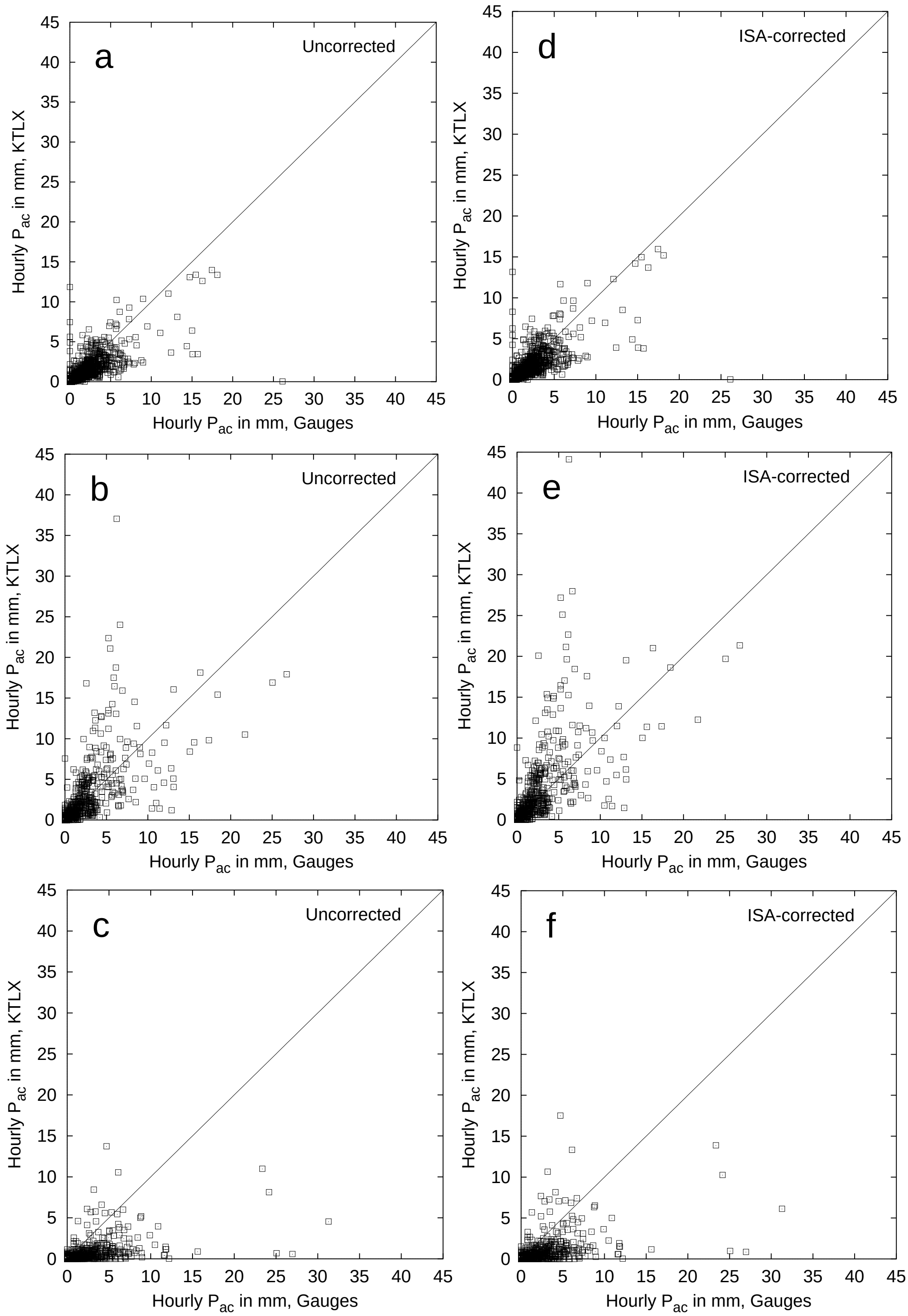

Fig. 4. Scatterplots of hourly precipitation accumulation for OK mesonet rain gauges and KTLX WSR–88D radar without (a–c) and with (d–f) ISA reflectivity corrections. Panels indicate range intervals from KTLX: a, d = up to 150 km, b, e = 150 to 200 km, c, f = 200 to 300 km.

corrected accumulation in the right polar plot. Radar–derived peak rainfall for this 48 h period was very large — the absolute maximum in the uncorrected case was 250 mm, and 300 mm with the ISA correction on Z, corresponding to a significant maximum of 50 mm in the difference field.

How well radar and gauge measurements compare is shown in the six scatterplots in Fig. 4. Uncorrected data is shown in the left panels (a)–(c), while the data with ISA correction are depicted in the right graphs (d)–(f). The data points give the hourly P_{ac} values measured by KTLX radar versus the corresponding measurements at the 114 available gauge sites (six sites are beyond KTLX 300 km range limit). In all panels of Fig. 4, the bulk of points is below 5 to 10 mm precipitation per one hour interval, and uncorrected peak values approach 40 mm. A general enhancement of the KTLX P_{ac} values by the ISA correction is noticeable in all right panels Fig. 4d–f.

Because QPE may be performed at varying ranges from radar, the subdivision in Fig. 4 has been chosen to show the range of the data points with respect to radar samples relative to the melting layer. Panels (a) and (d) denote ranges $r \leq 150$ km (52 gauges, 2311 points, below melting layer), panels (b) and (e) represent $150 < r \leq 200$ km (33 gauges, 1411 points, influenced by melting layer), and panels (c) and (f) are for $200 < r \leq 300$ km (29 gauges, 1309 points, above melting layer). We see that most of the points where the radar overestimates the surface rainfall, do indeed come from the range band where the radar beam intersected the melting layer. Panels (c) and (f) from above the melting layer show underestimation of surface rainfall by radar in general.

Application of the ISA Z–correction does not significantly affect the correlation coefficient or the standard deviation of the regression curve slopes in the scatterplots of Fig. 4. However, the mean values of the slopes in panels (d) and (e) become larger (0.63 in panel (d), compared to 0.57 in panel (a); 1.01 in panel (e) compared to 0.85 in panel (b)). The fact that the uncorrected slopes were well below unity in this case is not typical for KTLX. It might be due to the fact that only one standard Z–R relation was applied in this principal study. A more "stratiform" reflectivity–rain rate relation with coefficients $a = 200$, $b = 1.6$ would have led to a systematically larger precipitation accumulation.

4 Discussion

There are two main findings of this first observational study on the density stratification–related Z–correction according to Dotzek and Beheng (2001) and Dotzek and Fehr (2002). The first one is that it is indeed important to include this correction when computing precipitation sums from inversion of Z–R relations. For radar QPE ranges of up to 100 km, an enhancement of more than 10% in rainfall sums can be achieved. When larger ranges are also included, as in the USA or with the Nordic radar network NORDRAD, this can account to an up to 35% larger precipitation accumulation at 250 km range.

The second finding is that in order to really benefit from the Z–correction evaluated here, it is important to remove any artifacts from the measured reflectivity beforehand. The stratiform rain event described here had a strong melting layer signature. Application of the Z–correction factor further amplified this effect at some verification points. On the other hand, comparison of KTLX radar and Oklahoma mesonet rain gauges at points uncontaminated by the melting layer signature proved the use of the proposed correction method. A clear conclusion is that the density correction should be applied only to ranges where the beam is below the melting layer (in rain), as within the melting layer and above new Z–R relations and vertical velocities of ice particles hold. One could try to adjust R (by correcting Z) in the mixed phase and snow region, but that requires more assumptions than to make a VPR correction to the measured Z_e itself.

With proper ancillary information, one could know precisely where the radar is sampling below, within, and above the melting layer. With that information, one first needs to apply profile corrections to get a reasonable Z_e at the surface. Then, one could apply the density correction factor that would also change according to approximated particle fall speeds. One should be able to approximate those if there is good information available about the ranges at which the beam will intercept the melting layer. Finally, observed Z can be converted to R at the ground.

For this paper we have only used one horizontally homogeneous Z–R relation. In a more sophisticated step it would certainly be useful to more exploit the capabilities of the QPE SUMS system and to apply appropriate local Z–R relations depending if there is stratiform or (embedded) convective precipitation. However, our principal result that QPE algorithms at large range can be improved with the Z correction technique holds in any case.

Concerning the profiler–derived rapid decorrelation within average vertical reflectivity profiles, the caveat to further conclusions is the regional aspect of the presented measurements from orographically structured terrain. The observed low–level microphysical generation of precipitation in complex terrain may be extreme for this radar siting, and results may not be applicable to all other regions that receive stratiform rain. In any case, this statistical study suggests radar reflectivity measured at heights approximately 2 km above the radar are poorly correlated to surface reflectivity. The utility of reflectivity data measured at such heights and above, even if corrected using a VPR technique, are very limited for use in automated QPE algorithms.

5 Conclusions

Our study of air density effects on vertical profiles of R, and ways to eliminate them, has shown the following:

– The Z–correction procedure derived by Dotzek and Beheng (2001) leads to an increase in derived instantaneous rain rate (and therefore also in precipitation accumulation) of about 10% at 100 km, and almost 50% at 300 km range. For purposes of QPE, this is neither negligible nor minor, especially over complex terrain.

– Before applying the Z–correction procedure evaluated here, one has to make sure that artifacts in the Z field (like ground clutter and melting layer signatures) have been eliminated beforehand. Otherwise, unphysically high Z values would further be amplified. This supports the concept of operational VPR correction.

– Nevertheless, due to the strong decorrelation of vertical profiles of reflectivity, there is a maximum height and range for which application of the Z–correction is still soundly based. Without further assumptions, this limit is determined by the height of the melting layer.

Extension of the Z–correction above the melting layer after some VPR procedure remains an issue for future research.

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References

Beard, K. V., 1985: Simple altitude adjustments to raindrop velocities for Doppler radar analysis. J. Atmos. Oceanic Tech., 2, 468–471.

Dotzek, N., and K. D. Beheng, 2001: The influence of deep convective motions on the variability of Z–R relations. Atmos. Res., 59–60, 15–39.

Dotzek, N., and T. Fehr, 2002: Relation of precipitation rates at the ground and aloft — a model study. Revision submitted to J. Appl. Meteor..

Gourley, J. J., J. Zhang, C. M. Calvert, R. A. Maddox, and K. W. Howard, 2001: A real–time precipitation monitoring algorithm – Quantitative Precipitation Eestimation and Segregation Using Multiple Sensors (QPE SUMS). Preprints, Symp. on Precipitation Extremes: Prediction, Impacts, and Responses, Albuquerque, NM, U.S.A., Amer. Meteor. Soc., Boston, 57–60.

Gourley, J. J., R. A. Maddox, K. W. Howard, and D. W. Burgess, 2002: An exploratory multisensor technique for quantitative estimation of stratiform rainfall. J. Hydrometeor., 3, 166–180.

Gysi, H., R. Hannesen, and K. D. Beheng, 1997: A method for bright–band correction in horizontal rain intensity distributions. Preprints, 28th Conf. on Radar Meteor., Austin, TX, U.S.A., Amer. Meteor. Soc., Boston, 57–60.

Joss, J., and R. Lee, 1995: The application of radar–gauge comparisons to operational precipitation profile corrections. J. Appl. Meteor.., 34, 2612–2630.

Kitchen, M., and P. M. Jackson, 1993: Weather radar performance at long range — Simulated and observed. J. Appl. Meteor.., 32, 975–985.