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Comparison of quantitative snowfall estimates from weather radar, rain gauges and a numerical weather prediction model

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Quantitative snowfall estimates are needed in hydrology and weather services. Snow measurements with weather radars and rain gauges are challenging, hence adjustment methodologies have been developed for them. Still, the accumulations from different data sources differ. We compared different data sets from radars, gauges and numerical weather prediction models, commonly used in operational or semi-operational applications with varying corrections applied. The mean ratio of radar-based quantitative precipitation estimates (QPE) to the reference data sets varied between 0.43 and 1.59. Variability of the microphysical properties of snow is so large that a perfect operational solution for all situations may be impossible to reach. However, much improvement could be achieved if the presently-known methods were applied to all snowfall measurements.

Introduction

In hydrology, radar has been used for measuring rainfall. So far, less attention has been paid to snowfall. To estimate the amount of melting snow, accurate and especially unbiased measurements of the amount snowfall are needed.

Quantitative snowfall estimates are also needed for numerical weather precipitation (NWP) models, which use the data for verification and assimilation. The spatial and temporal resolution of radars would make them a very attractive data source for NWP, if the data quality is sufficient. Accurate, high-resolution, quantitative snow data may also be required for avalanche forecasting and road maintenance in winter.

Typically, radar measurements of reflectivity (Z) are converted into rainfall intensity (R) using empirical equations of the form of a power law $Z = aR^b$, where parameters a and b are usually determined from a large set of measurements: currently a = 200 and b = 1.6 are widely used (Marshall *et al*. 1955). Even though the equation

$$Z = 200R^{1.6} \quad (1.1)$$

was developed in the United States for rain, it is operationally used for all precipitation, e.g., in the European radar project OPERA, and it was used for comparison also in this study. For low reflectivities, it gives weaker intensities than all the dedicated snow equations.

Battan (1973) published a review of 69 other pairs of the a and b values valid for different climates and precipitation types. The variability of the power law parameters is related to drop size distribution, as Z is proportional to the 6th power

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of the particle diameter (D), while R is proportional to particle volume and fall speed, the combined effect of which can be approximated by the 3.7-order moment of D.

For snowfall, fewer equations have been published, and the ranges of the parameters a and b are expected to be even greater than for rainfall, as the size and shape of snow particles is more diverse than raindrops.

The Finnish Meteorological Institute (FMI) uses different equations for rain (from Dölling *et al*. 1998):

$$Z = 316R^{1.5} \quad 1.2$$

and snow [based on Sekhon and Srivastava (1970)]

$$Z_e = 398S^{2.21} \quad 1.3$$

to convert radar reflectivity, Z, into precipitation intensity (R or S, mm h^{-1}). Z_e is the equivalent radar reflectivity factor, which differs from the cloud physical definition of Z, because the dielectric constant ε has in the radar signal processing the value referring to water and not for ice. It is usually expressed as $|K|^2 = |(\varepsilon - 1)/(\varepsilon + 2)|^2$, for water $|K|^2 = 0.93$ and for ice $|K|^2 = 0.2$)

Equation 1.3 leads to significantly smaller intensities than any of the other snowfall equations, and even smaller than Eq. 1.1 for reflectivities above 15 dBZ.

Equation 1.3 was used until 2004. However, the end users reported an underestimation of snowfall, hence the equation

$$Z_e = 100S^2 \quad 1.4$$

was therefore implemented (Saltikoff *et al*. 2010). We also tested a quite similar equation

$$Z_e = 75S^2 \quad 1.5$$

used for the NEXt Generation Radar network (NEXRAD) in the United States by Zhang *et al*. (2014).

Usually, the quality of different $Z(R)$ equations is evaluated against gauge measurements. Precipitation measured with manual gauges is almost always smaller than the actual one and, especially for snowfall, the aerodynamic error is the largest source of uncertainty. The Finnish Environment Institute (SYKE) applies an operational procedure for the daily precipitation observation correction of aerodynamic, wetting and evaporation errors. The aerodynamic correction method used has been developed for Finnish conditions in which wind observations at the gauge are not available but the data from the nearest synoptic station can be used (Solantie 1986).

In many studies, quantitative precipitation estimates (QPE) of snowfall are either not mentioned (e.g. Tabary 2007, Germann *et al*. 2006), deliberately excluded (e.g. Lopez 2011), excluded by limiting the studied period to summer months (e.g. Szturc *et al*. 2011) or are isolated case studies (e.g. Huang *et al*. 2014). However, in operational applications, solutions which can be applied every day and everywhere are needed. Existing solutions are known to be suboptimal and the users should be made aware of the magnitude and sources of the inevitable errors in the data.

This study is a follow-up to Saltikoff *et al*. (2010), which described the operational radar network of Finland. In that study processing radar data for snowfall was mentioned without any quality estimates as no reliable reference data were available at the time of writing that paper. In this paper, we compare the radar-based snowfall estimates with (i) corrected 24-h gauge measurements from Finland, (ii) uncorrected 6-h gauge measurements from Europe, and (iii) a first-guess field produced by a NWP model. We report improvements in quality control of radar and gauge data, and we attempt to assess the error structure of remaining uncertainties in these data sources.

Known error sources

Radar error sources

The uncertainty factors affecting radar reflectivity are electronic miscalibration, beam blocking, and attenuation due to both precipitation (Battan 1973) and wet radome (Germann 1999). The vertical profile of reflectivity (VPR) has been mentioned as the greatest source of uncertainty in radar measurements at high latitudes, which leads to a range-dependent error (Zawadzki 1984). At large distances, the radar probes the upper parts of the cloud, where reflectivity is weaker. Also the radar beam gets wider with increasing distance, which more easily leads to partial beam filling either near the cloud top or near the edges of cloud area. Many countries have developed and implemented correction methods for VPR, which also compensate for overestimation related to large, partially-melted snowflakes in the melting layer (e.g. Koistinen *et al*. 2004). Instead of studying the vertical structure near the radar, the compensation of range-dependent errors can be developed as an approach using gauges or satellites as the "ground truth" (e.g. Gabella *et al*. 2011) As the errors related to the vertical profile of reflectivity are greatest at long distances from a radar, dense radar networks suffer less from this phenomenon.

Total beam overshooting occurs when the precipitation top does not extend to the height of the lowest radar measurement. This is typically associated with drizzle and snowfall (especially in a cold climate). There is no obvious solution to this problem, apart from increasing radar network density. On the European scale, the error only becomes substantial at the edges of the image or in areas with low radar density (e.g. eastern Finland, Ireland) or complex topography (e.g. Alps, Norway). In central Europe, measurements collected far from a given radar can usually be replaced with data from a nearer radar.

The concept of using one equation to convert radar reflectivities into precipitation intensities relies on the assumption that the particle size distribution is constant. Everyday experience shows, however, this is not the case especially not for snowfall; indeed, it has even been claimed that there are no two similar snowflakes. Huang *et al*. (2014) used a 2D-video disdrometer to derive $Z_e(S)$ relations in four cases and found that different snow types lead to different optimal $Z(S)$ relations. In practice, the challenge lies in the determination of snow type. Unfortunately, optical disdrometers are research instruments, which cannot be widely used. Even if such observations were available locally, their representativeness would remain limited in space and time. Great variability of snowflake types has been demonstrated by Tollman *et al*. (2008) who compared particle density to observations of particle type during a warm front overpass and found several different classes of snowflakes within a few hours.

Gauge error sources

Gauge precipitation measurements are always subject to various sources of errors and therefore the measured precipitation is usually lower than the actual one. Sevruk (1986) classified the errors as follows: (1) random and systematic errors of point precipitation measurements, (2) random error at the gauge site (due to local irregularities of topography and micro-climatic variations), and (3) random error of a gauge network (due to inadequate network density). A systematic error is considered the most influential source and can be further divided into several components: (1) aerodynamic, (2) wetting, (3) evaporation, (4) splash in and out, and (5) blowing and drifting snow. According to Førland *et al*. (1996), the most important of these for the Nordic countries are the first three. Especially for snowfall, the aerodynamic error is the greatest and most difficult to eliminate. It depends on air flow conditions around the gauge, gauge-top height above ground, wind speed and direction and precipitation type. One should note that gauge height is not at all standardized worldwide (e.g. Lopez 2013).

In northern Europe, the errors related to undercatch of snow are considerable. For instance, 5.2% of the annual precipitation in Germany occurs as snow and 13.2% as a mixture of rain and snow. Errors due to wind-induced undercatch are estimated to be 12.3% ± 3.1% during winter, and to 5.6% ± 1.7% during summer. In a long-term average, the total error has been estimated as 16.7% (Richter 1995, Paulat *et al*. 2008). In Finland, the long term (1961–2011) corrected average annual precipitation is 674 mm, in which proportions for liquid and solid precipitation forms are 65.5% and 34.5%, respectively, if a mixed form is assumed to be 50% liquid and 50% solid. The mean correction factor (corrected precipitation/observed precipitation) for all forms of precipitation is 14.1% and the median correction factors are

1.10, 1.22 and 1.40 for liquid, mixed and solid precipitation on the daily basis, respectively.

When data are used for hydrological purposes, it is usually corrected at least for wind-induced and evaporation errors. However, this is usually not the case for real-time measurements sent through the Global Telecommunication System (GTS), which are typically used in NWP applications (e.g. for nowcasting, model validation or data assimilation).

The World Meteorological Organization (WMO) regularly organizes solid precipitation measurement intercomparison campaigns to estimate the errors for different gauge types. The latest one of such intercomparisons is the ongoing SPICE project (Nitu *et al*. 2012). The earlier intercomparisons have shown, that in low temperatures (when snowflakes are light and fluffy) and strong winds, the correction factor for Tretjakov's gauge can reach up to 3, which means that gauges may sometimes collect only around 30% of the actual snow amount (Goodison *et al*. 1998).

Measurements by many types of automatic gauges currently in use can be affected in different ways. Heated tipping buckets are used in areas with short winters, but they are prone to loss through evaporation. Automatic weighing gauges are an alternative in countries where the season with freezing temperatures is long. One serious operational problem with weighing gauges is that wet snow can stick to the gauge. Other problems include gauges catching drifting snow and wind-induced oscillation of the weighing mechanism (wind pumping) (Goodison *et al*. 1998)

Optical instruments, mainly developed as present weather sensors, can also be used for quantitative precipitation estimates. Wong (2012) compared weighing gauges and optical instruments in Canada. In his study, one of the two optical instruments showed an overcatch, the other one an undercatch. The overcatch got worse with increasing wind speed, which clearly indicates that the error structure for optical instruments is different from that of manual or weighing gauges.

Representativity differences

A gauge measures a small volume near the ground, a radar measures large volume aloft. Hence, when comparing radars and gauges, an additional challenge arises from the different sample sizes of these instruments.

Radar measurement volume can be hundreds of meters thick and several kilometres wide and high (1° beam is ca. 3 km wide at a 175 km range), while the measurement area of a gauge is typically 400 cm^2 (gauge orifice) or 100 cm^3 (in volume, for an optical instrument measuring scattering).

At longer distances, radar measures snow at rather high altitude. After the snowflakes have been measured, they may drift significant distances with the wind before reaching ground level. This is more serious for snow than for rain, because snowflakes fall with smaller fall speeds. Hence the error produced by wind drift can cause substantial differences between QPE from a radar pixel and co-located gauge measurements (Mittermaier *et al*. 2004).

What is snow and what is rain?

Precipitation phase (snow or rain) is essential for radar QPE, important for gauge QPE and relatively difficult to observe or forecast.

In the radar world, we have to distinguish between the phase of measured hydrometeors and the phase of hydrometeor at ground level. Very often, a radar would detect snow that will melt further down and be collected as rain by a gauge.

The phase of measured hydrometeors can be estimated from the shape and structure of the falling particles as determined using dual polarization radars. These radars send both horizontally and vertically polarized microwaves. The ratio of the two measured reflectivities is related to the shape of the measured particle. The pulse-to-pulse correlation can be used to estimate the uniformity of the fall speeds of the particles, which in turn can be used to identify the presence of partially melted snow. Most operational hydrometeor classification schemes also use the height of the freezing level as an input parameter.

In early numerical weather prediction applications, the precipitation phase determination was based on relative topography (geopotential height between 1000 hPa and 700 hPa surfaces), which strongly correlates with average temperature in the column between these two surfaces (e.g. Heppner 1992).

In many limited-area models today, different hydrometeors are forecast parameters and the water phase can be derived from them for those grid points where the model predicts precipitation. Of course, some uncertainties can affect model predictions of hydrometeors as important assumptions need to be made about the size distribution and physical properties of the latter. If model and radar on the location of precipitation, the model parameter cannot be used directly. Hence, many operational applications in weather and hydrological services use a method based on surface temperature and humidity (Gjertsen and Odegaard 2005, Saltikoff *et al*. 2010).

At the FMI, the form of precipitation is estimated from temperature and humidity using the following equation developed by Koistinen *et al*. (2004):

$$P_{\mathrm{lp}} = \frac{1}{1+e^{22-2.7T-0.2H}}, \qquad (2)^*$$

where P_{lp} is the probability of liquid precipitation, T (°C) is the temperature, and H (%) is the humidity at a height of 2 m. If $P_{\mathrm{lp}} < 0.2$, precipitation is considered solid and Eq. 1.4. is used for radar-based QPE. If $P_{\mathrm{lp}} > 0.8$, precipitation is considered liquid and Eq. 1.2 is used for QPE. In the case of $0.2 \le P_{\mathrm{lp}} \le 0.8$ a weighted combination of these two is used.

In the global model of the European Centre of Medium Range Forecasting (ECMWF), rain and snow amounts are treated as prognostic variables and the melting of snow (kg kg^{-1} s^{-1}) at each model level is parameterized as

$$M_{\mathrm{snow}} = a_{\mathrm{P}} \frac{c_{\mathrm{P}}}{L_{\mathrm{f}}} \frac{\left(T_{\mathrm{w}} - T_{\mathrm{melt}}\right)}{\tau}, \qquad (3)$$

where T_{w} is the wet-bulb temperature, T_{melt} is the melting temperature (namely 273.15 K) and a_{P} is the fraction of the grid box covered by precipitation. Constants c_{P} and L_{f} are the specific heat at constant pressure for dry air and the latent heat of fusion of water, respectively. The time scale τ (in seconds) is defined as

$$\tau = \frac{7200}{1+0.5\left(T_{\mathrm{w}} - T_{\mathrm{melt}}\right)}, \qquad (4)$$

This formulation ensures that precipitation can remain frozen at temperatures up to about +5 °C, particularly when the air is very dry.

Data and methodology

Radar *versus* individual gauges

The radar data were compared with 24-hour snowfall from the FMI gauge stations. The period of the study was December 2010 to March 2011, and snow was observed on 76 days.

At that time, the FMI radar network consisted of 8 C-band Doppler radars, three of which (VAN, IKA and ANJ) had dual-polarization capability (Fig. 1). The radar data were processed by using a dual polarization-based filter to remove non-meteorological echoes. Thus, only the three dual-polarization radars were used in this study.

The radar data were collected every 5 minutes. Only the second lowest elevation angle, 0.7°, was used, as the lowest one may suffer from partial beam blocking. The effect of measurement height was compensated by applying a correction for VPR (Koistinen and Pohjola, 2014). Then the reflectivities were converted to intensities using Eq. 1.4 (the "FMI snow equation") and then aggregated into 24-h sums. In case of missing data, the entire 24-h period for that station was excluded from the data set. Each radar was treated separately, so that a gauge located between two radars could be a part of two radar-gauge pairs. In total, 5318 observation pairs were processed. Only pairs in which both radar and gauge exceeded 0.25 mm were used.

The gauge data were corrected for aerodynamic, wetting and evaporation errors. The common expression for corrected precipitation P_{c} is written as

$$P_{\mathrm{c}} = (1 + A)(P_{\mathrm{m}} + \Delta P_{\mathrm{w}} + \Delta P_{\mathrm{e}}), \qquad (3)$$

where $1 + A$ is the aerodynamic correction, P_{m} is the measured precipitation, ΔP_{w} is the wetting correction and ΔP_{e} is the evaporation correction.

The aerodynamic correction procedure, called the exposure method, was specially developed by Solantie (1986) for conditions when

* On 28 September 2016 the equation was corrected in the following way: 2.2*T* (in the denominator) was replaced by 2.7*T*.

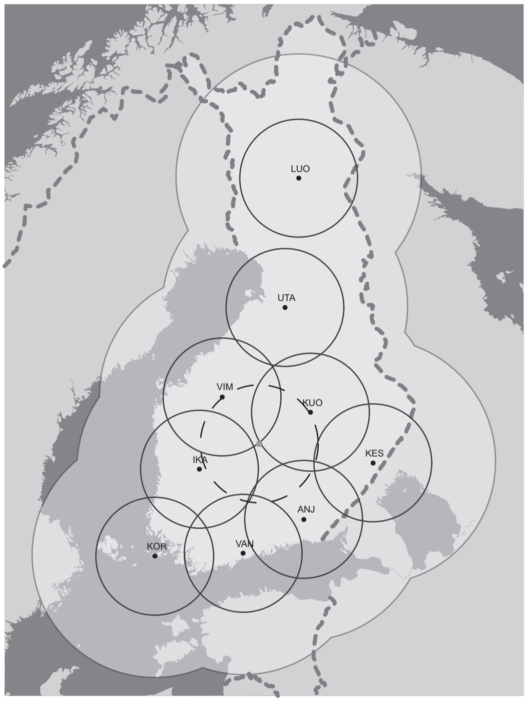

Fig. 1. Position of the FMI radars in Finland. The combined maximum coverage range of 250 km is the outer line, and distance of 120 km is indicated with a black circle around each radar site. During the time of this study, only VAN, IKA and ANJ were dual-polarization radars and hence they were the only ones used in this study.

wind speed observations at the gauge are not available, but the weather data from the nearest synoptic station representative for the particular area can be used instead. Factor A depends on wind speed and direction, precipitation type, fall speed and gauge type. It also accounts for the sheltering by obstacles surrounding the gauge in the wind direction during the storm. The sheltering effect is estimated as a function of the height angle, i.e. the height of the obstacle in relation to the distance of the obstacle to the gauge.

The values of both ΔP_w and ΔP_e used in this application were obtained from the recommendations of Førland *et al*. (1996). ΔP_w depends on the material of the gauge and the form of precipitation and is inversely proportional to the orifice area of the gauge. ΔP_e varies with gauge type and month of the year.

Scandinavian radar, gauge and model data sets

The radar composites over Scandinavia were compared with both gauge observations and first-guess fields of numerical weather prediction model for the period Dec. 2012–Feb. 2013. The radar composites came from the data hub of the European radar project OPERA. They are available every 15 min and with a spatial resolution of 2 km (Huuskonen *et al*. 2014). These data were accumulated over 6-hour periods (ending at 00:00 UTC, 06:00 UTC, 12:00 UTC and 18:00 UTC) prior to computations. Only pixels that were flagged as valid precipitation data (i.e. not flagged as 'no data' or 'undetected') were retained in the accumulations. In the OPERA rain rate product, the VPR correction described in the previous paragraph was not implemented, and the standard $Z(R)$ relationship of Eq. 1.1 was always used. For this study, the snowfall intensities were recalculated using an inversion method: in pixels identified as snow, R in the OPERA rain rate product was converted back to reflectivity (Z_e) using Eq. 1.1, and then the reflectivity was converted to snowfall intensity using Eqs. 1.4. and 1.5.

The second precipitation data set used in this study were 6-hourly accumulations from synoptic stations via the Global Telecommunication System (GTS) of the WMO. These data were taken from the GTS as such, so the correction described in the previous paragraph was not applied. Note that not all countries provide 6-hourly gauge measurements on the GTS.

The third precipitation data set consisted of short-range forecasts obtained from the operational forecasting system of the ECMWF, which is described in Courtier *et al*. (1994). The comparison of the radar with model data was conducted for 6-hour periods to be consistent with the validation against SYNOP rain gauges that was performed for 6-hour accumulation periods ending at 00:00, 06:00, 12:00 and 18:00 UTC. Therefore, to sample the four 6-hourly accumulation periods contained in each day, 6, 12, 18 and 24-h-range forecasts (starting at 00:00 UTC) were needed. For instance, the 6-hour accumulation period ending at 12:00 UTC would be obtained as the difference between the 12:00 and 06:00 forecasts.

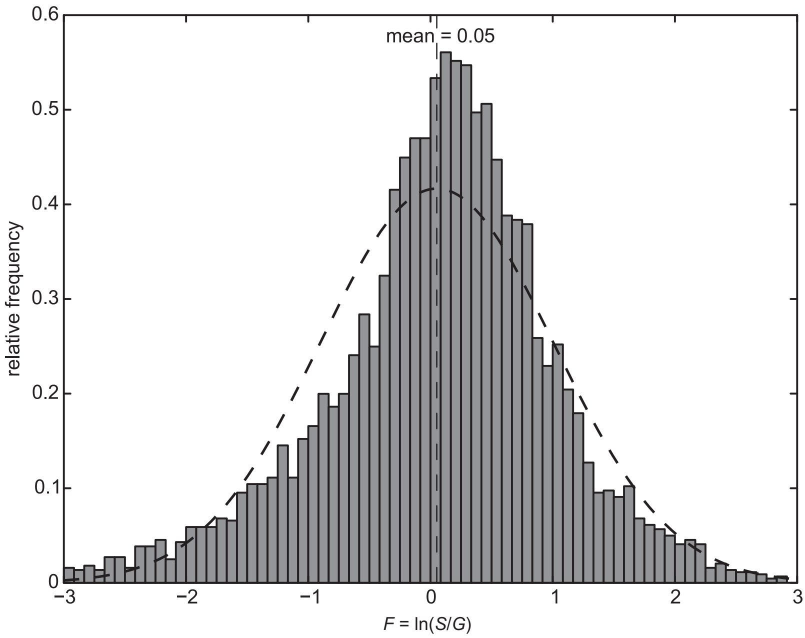

Fig. 2. Distribution, mean and variance of logarithmic radar/gauge ratio F for snowfall in Finland in winter 2010–2011. S is the 24 h accumulated snowfall (mm) from radar measurements, and G is the 24 h accumulated snowfall (mm) from gauge measurements.

The forecast data were produced at the operational horizontal spectral resolution of T1279 (i.e. roughly 16 km) and with 137 vertical levels. In the comparison of radar with model data, precipitation amounts at OPERA 2-km pixels were averaged over each ECMWF model grid box (≈ 16 km) to avoid spatial representativeness issues. Even though model forecasts are likely to be usually less accurate than SYNOP observations, they are used here in order to assess the differences found between radars and gauges. They also have a better spatial coverage than 6-hourly gauge data, even though the spatial resolution is worse.

The area covered by our 'Scandinavian data set' also includes most of Finland and Estonia and is between 6°–30°E and 55°–72°N. Lopez (2014) considered nine other European geographical subdomains. The Alpine region would also be of interest for snow studies, but this is not an option as long as radars in Italy, Austria and Switzerland are not included in the OPERA composites.

Results

Radar *versus* individual gauges

The radar-based estimates of 24-h precipitation were in general larger than gauge measurements (Fig. 2). The average radar-to-gauge ratio for the entire data set was 1.59 (Table 1). The differences depended on precipitation intensity: the weaker the intensity was, the greater the difference in the positive direction. This was most common for precipitation below 2 mm day^{-1}. On the other hand, in moderate and high-intensity cases, one could find cases in which radar underestimated precipitation (Fig. 3). The radar-to-gauge ratio varied with the distance to the radar (Fig. 4). At shorter ranges radar gives typically greater values than gauge. Beyond 150 km, the radar gives increasingly smaller values.

Table 1. Results of all the comparisons.

Reference data set	Radar equation	Mean radar/ gauge ratio
24-h individual gauges	1.4	1.59
Scandinavia, gauges	1.1	0.43
Scandinavia, model	1.1	0.38
Scandinavia, gauges	1.4	0.83
Scandinavia, model	1.4	0.72
Scandinavia, gauges	1.5	0.95
Scandinavia, model	1.5	0.84

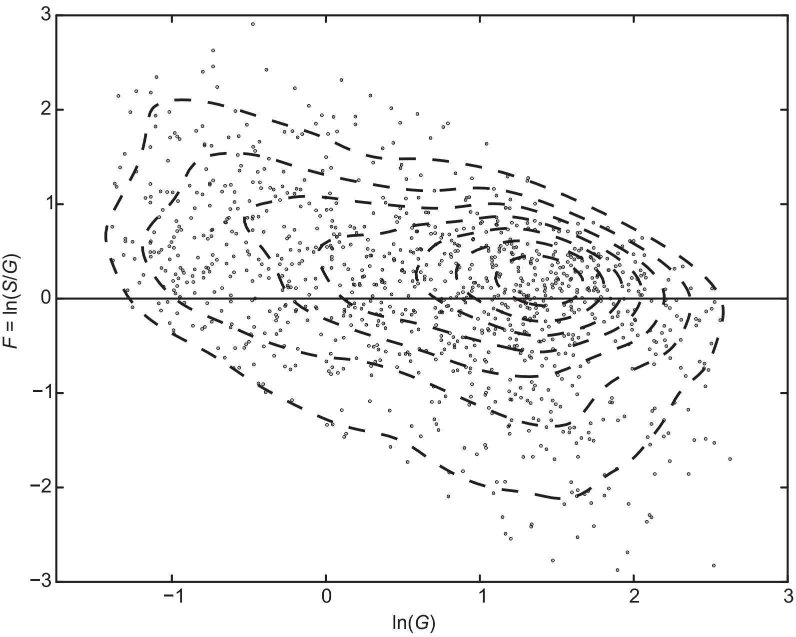

Fig. 3. Logarithmic radar/gauge ratio F as a function of the gauge measurements (G) of 24-h snowfall in Finland in winter 2010–2011 (–1 corresponds roughly to 0.35 mm, 1 to 3 mm and 2 to 7.5 mm). Dashed lines are isolines of density of radar–gauge pairs in the scatter plot.

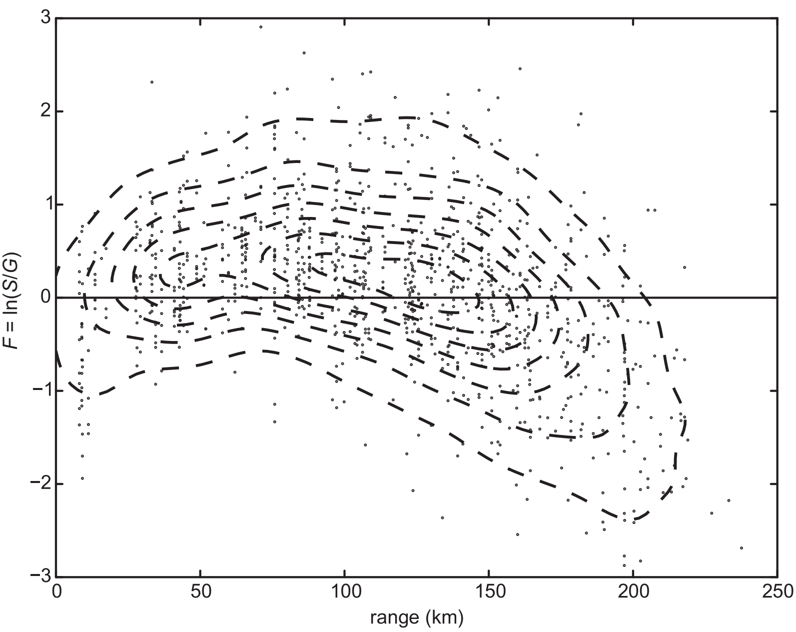

Fig. 4. Logarithmic radar/gauge ratio F as function of distance from the radar for snowfall in Finland winter 2010–2011.

Scandinavian radar, gauge and model data sets

In the second experiment, the mean precipitation rate over all Scandinavia was compared with the model values and gauges using only the pixels where both data sources were available. The main difference between the results from the two experiments was that for the Scandinavian data set the radar-based estimates were smaller than reference values (Table 1). Time series illustrates the well-known underestimation of snow when Eq. 1.1 was used. Radar-based QPE was on average 43% of that of the gauges' and 38% of that of the model's (Fig. 5a and b). When Eq. 1.4 was used to process the radar data, the mean values were closer to both gauges (83%) and model (72%) (Fig. 5c and d), but underestima-

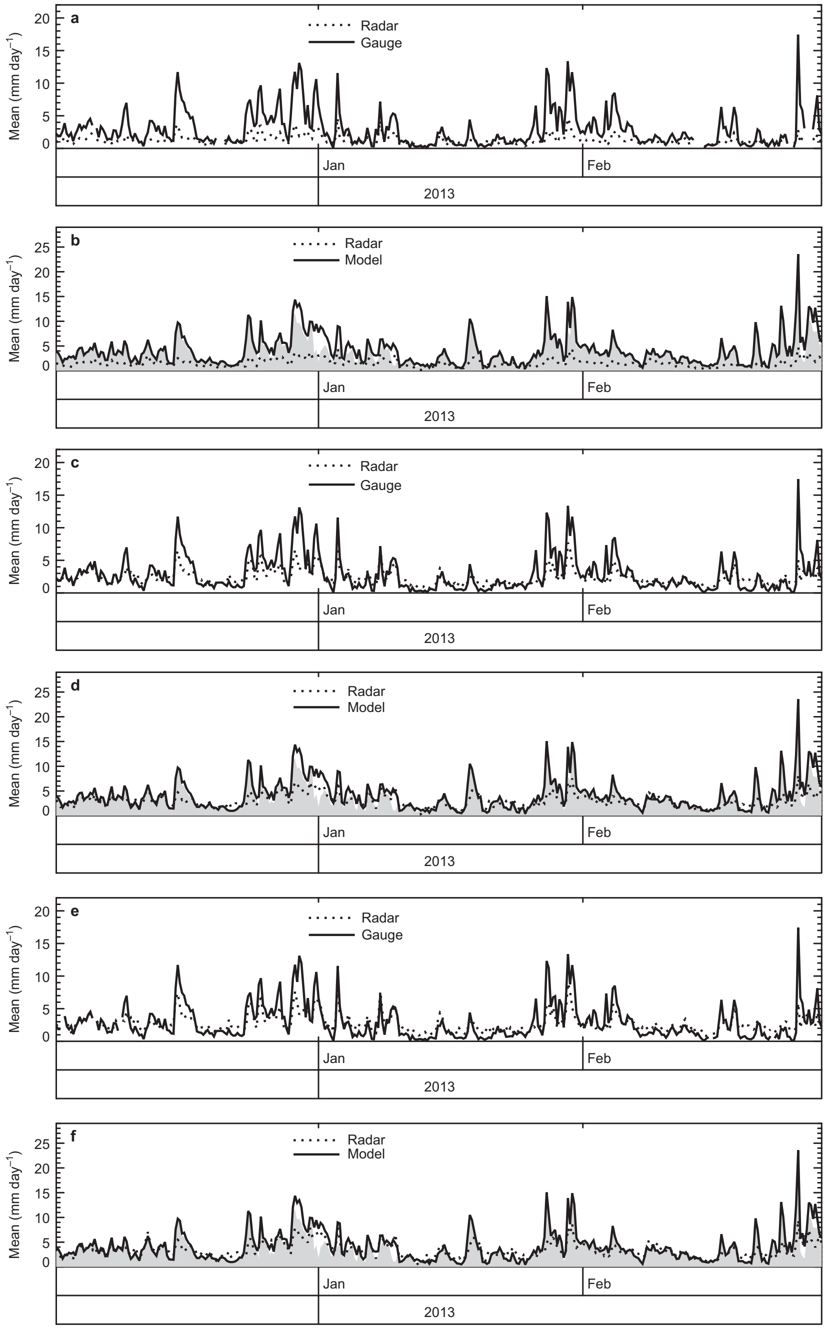

Fig. 5. Time series of 6-hour precipitation accumulations averaged over Scandinavia between 1 December 2012 and 28 February 2013. The grey areas in panels **b**, **d**, and **f** show the amount of snowfall from the ECMWF model, while solid lines represent total precipitation (both rain and snow). (**a**) Radar using equation $Z = 200R^{1.6}$ (standard OPERA product), gauges from GTS (means = 1.26 and 2.92 mm day^{-1}, respectively). (**b**) Radar using equation $Z = 200R^{1.6}$ (standard OPERA product) and ECMWF model first guess (means = 1.55 and 4.13 mm day^{-1}, respectively). (**c**) Radar using $Z_e = 100S^2$ (FMI equation) and gauge from GTS (means = 2.38 and 2.88 mm day^{-1}, respectively). (**d**) Radar using $Z_e = 100S^2$ (FMI equation) and ECMWF model first guess (means = 2.99 and 4.13 mm day^{-1}, respectively). (**e**) Radar using $Z_e = 75S^2$ (NEXRAD equation) and gauge from GTS (means = 2.75 and 2.88 mm day^{-1}, respectively). (**f**) Radar using $Z_e = 75S^2$ (NEXRAD equation) and ECMWF model first guess (means = 3.45 and 4.13 mm day^{-1}, respectively).

tion of high snowfall intensities was still visible. When Eq. 1.5 was used for radar, the mean values were even closer to gauges (95%) and model (84%). However, Eq. 1.5 also tended to overestimate smaller snowfall amounts (Fig. 5e and f).

Conclusions and prospects

Compared with 24-h gauge measurements, the radar overestimated snowfall up to the distance of 150 km, especially at light and moderate intensities. In comparison with the Scandinavian data set, radar QPE was on average smaller than the gauge or the model estimates. The major differences between the two data sets were that in the Scandinavian data set the radar data were not corrected for vertical profile of reflectivity, and the aerodynamic, wetting and evaporation correction was not applied to gauge measurements.

When radar gives systematically greater values than gauge, the reason can be overestimation by the radar or underestimation by the gauge. This happens for low precipitation intensities and at short distances from the radar. There are no known reasons which would cause a gauge to show too large values, so if the radar shows smaller values than a gauge, either the radar is underestimating or our correction method for gauges is over-compensating errors. The VPR correction compensates for some of the range-dependent errors in the radar data, but because snow sometimes falls from rather shallow clouds, the radar beam may rise over the entire precipitation system. Compensation with VPR is not possible for such cases.

The effect of wind drift is large when short measurement periods are considered, but here the 6-h or 24-h accumulation period and, in the case of the model, the 16 km^2 integration area compensated the effect of wind drift.

Several weather services and hydrological services combine operationally the radar and gauge measurements to get the accuracy of the point measurement and resolution of the radar. Some techniques for this are described by Gabella *et al*. (2001) and Michelson and Koistinen (2000). In this work, the difference of radar and gauge measurement was strongly affected by the use of (i) aerodynamic, wetting and evaporation correction to the gauges, (ii) vertical profile correction to radar and (iii) selection of the *Z–S*. For shorter accumulation periods, the effect of the wind drift would become important, and probably also an advanced method for aerodynamic correction would be needed. We found large uncertainties in both radar and gauge data, hence they should be combined with care.

For the NWP applications in which model and radar observations need to be compared (e.g. model validation, data assimilation), an alternative approach is to compute simulated reflectivities from model output fields and compare those with observed radar reflectivities, rather than to try to retrieve precipitation from the radar observations. Employing such forward operator has the potential to ensure a better consistency with the microphysical assumptions already made in the model physical parameterizations, provided these are realistic enough. This is the method used in many limited area models (e.g. Wattrelot *et al*. 2014). On the other hand, ECMWF is aiming to assimilate OPERA two-dimensional precipitation composites in their operational, global, variational data assimilation system in a similar way as what is already done with NCEP Stage IV radar and gauge composites for the USA (Lopez 2011). However, assimilation of the European radar data will not become operational before quality of radar-based QPE has been improved especially for snowfall.

Works by Tollman (2008) and Goodison (1998) showed that there is a multitude of snow types which can rapidly vary in time depending partly on temperature. Hence, the use of different *Z–S* equations for different meteorological situations would be beneficial. Two challenges remain: (1) to derive such equations, and (2) to determine when to apply each of them. Dual polarization radars are expected to help as they provide additional information about the properties of snow flakes. They, however, will not resolve all the issues related to quantitative assessment of all the different kinds of snow.

For the applications described in this work, improvements are to be expected. The FMI and SYKE are improving the correction procedure. To develop an improved version of Eq. 1.4, a larger data set is needed. At the European level, the OPERA team is currently working on the implementation of a correction procedure for the vertical profile of reflectivity in their European composites. Such procedure is already in use at the FMI and many other national weather services.

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