Mapping Radar Reflectivity Values of Snowfall Between Frequency Bands

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Abstract—Motivated by the use of a $K_u/K_a$-band radar in the Global Precipitation Measurement core satellite due to launch in 2013, we have studied the use of techniques to simulate radar observations of snowfall at these bands from data at C/W-bands by using scattering simulations to derive the approximate relationships of radar observables at the C/W-bands and those at $K_u/K_a$-bands. In this paper, we form the cross-band relationships by simulating radar reflectivity and attenuation in snowfall. The relations can be used to simulate observations at given bands from measurements at other bands. When measurements are available at multiple frequencies, the consistency of the model and the measurements can be used as a measure of the validity of the underlying assumptions of snowflake shape and constitution, which have been an important topic in recent snowfall remote sensing research. We also present mapping functions that can be used to perform cross-band analysis of snowfall observations and examine the use of these for practical cases of combined ground- and space-based (CloudSat) radar measurements as well as airborne radar data from the 2003 Wakasa Bay experiment.

Index Terms—Radar reflectivity, snowfall, weather radar.

I. INTRODUCTION

PRECIPITATION is one of the critical components of the Earth’s hydrological cycle and is one of the Essential Climate Variables affecting the climate and its change. Due to the sparsity of in situ observation networks, spaceborne measurements are currently the only technologically feasible method to implement precipitation cycle monitoring at a global scale. With the recent growth in the interest in observing the Earth’s climate, the motivation to use precipitation radars on satellites has likewise increased.

Two missions equipped with precipitation radars are currently on Earth orbit: the Tropical Rainfall Measurement Mission (TRMM) and CloudSat satellites, both operated by the National Aeronautics and Space Administration (NASA). Additionally, the Global Precipitation Measurement (GPM) core satellite codeveloped by NASA and the Japan Aerospace Exploration Agency, and the Earth, Clouds, Aerosols and Radiation Explorer (EarthCARE) developed by the European Space Agency are in advanced stages of preparation. The TRMM uses a single precipitation radar at 13.8 GHz ($K_u$-band) [1]; in GPM, a more sensitive radar at the same frequency is used together with an additional 35.6-GHz ($K_a$-band) radar [2], [3]. CloudSat [4] and EarthCARE [5] take a slightly different approach, utilizing far more sensitive but attenuation-prone 94-GHz (W-band) radars that can measure both clouds and precipitation.

Ground-validation activities for GPM, due to launch in 2013, are currently undertaken by NASA and various international partners to characterize the expected GPM observations in various geographical regions [6], [7]. Finland is an active member in the NASA Precipitation Measurement Missions science team, specializing in GPM high-latitude ground validation. The Finnish Meteorological Institute operational radar network and additional research radars operated by the University of Helsinki and Vaisala provide a dense and nearly complete coverage of Finland, located at the northern extreme of the GPM orbit (inclination 65°) between latitudes of approximately 60° N and 70° N. At these latitudes, snow is an important form of precipitation even at sea level, requiring special attention in the ground validation. Overall, 35% (in terms of water-equivalent total precipitation) of surface precipitation in Finland is snow [8]; it is the dominant type at the surface for four to eight months (depending on the location) in the winter and is observed at higher altitudes throughout the year. Finland’s location at the coast of the Baltic Sea, west of the vast Eurasian landmass, ensures that both oceanic and continental types of winter precipitation are experienced within the country.

As a part of the GPM ground-validation effort, the Light Precipitation Validation Experiment took place in the region around Helsinki, Finland, in the fall of 2010 [9], making use of the Helsinki Testbed mesoscale network. In support of the experiment, as well as further activities before and during the GPM mission, we have developed multifrequency methods that can be used to simulate radar measurements of snowfall at the $K_u$- and $K_a$-bands from data at C- and W-bands. Such simulation techniques are required to generate expected GPM radar profiles for the region, as the Finnish radar network is made up entirely of C-band radars. As a source of the C-band data in the current study, we use the Finnish ground-based radars, particularly the University of Helsinki radar in Kumpula. For
TABLE I

| Band | Frequency | $|K_w|^2$ | Geometry               |
|------|-----------|----------|------------------------|
| C    | 5.6 GHz   | 0.93     | Horizontal (ground)    |
| Ku   | 13.6 GHz  | 0.93     | Vertical (air/space nadir looking) |
| Ka   | 35.6 GHz  | 0.92     | Vertical (air/space nadir looking) |
| W    | 94.0 GHz  | 0.75     | Vertical (air/space nadir looking) |

W-band, the main source is the CloudSat spaceborne Cloud Penetrating Radar.

II. SNOW SIMULATIONS AND FREQUENCY MAPPING

A. Snowfall Model

To develop a model to predict radar reflectivity properties of snow at one wavelength, given those at another, we have constructed a method to simulate scattering from snowfall profiles. In this setup, reflectivity $Z_{h,v}$ ($h$ and $v$ refer to horizontal and vertical polarizations, respectively) is computed by integrating the computed radar cross section over the particle size distribution as

$$Z_{h,v} = \frac{\lambda^4}{\pi^3 |K_w|^2} \int_0^{D_{\text{max}}} \sigma_{h,v}(D) N(D) dD \quad (1)$$

where $\lambda$ is the wavelength, $K_w = (m_w^2 - 1)/(m_w^2 + 2)$ is the dielectric factor of water at the wavelength, computed from the complex refractive index of water $m_w$, $\sigma_{h,v}(D)$ is the radar cross section for particle diameter $D$, and $N(D)$ is the size distribution function. The reflectivity in units of dBZ is obtained by taking $10 \log_{10}(Z)$ with $Z$ in units of mm$^6 \cdot$ m$^{-3}$. For the nonspherical snow particles in this paper, we always define $D$ as the diameter of equivalent volume spheres, $D_{\text{max}}$ being the largest such diameter to appear in the particle ensemble. Horizontal polarization is used for all results.

The radar cross section for individual particles is computed using a freely available T-matrix scattering simulation code [10], [11], modified slightly to interface to analysis tools. This code can simulate nonspherical particles with a free selection of incidence and scattering angles for radiation as well as particle orientation and user-definable particle size, wavelength, and complex refractive index. Snow particles are modeled as partially aligned oblate spheroids with an aspect ratio $r$ and a normally distributed canting angle $\beta$ with a mean of zero and a standard deviation $\sigma_\beta$. The measurement geometry is selected to be either horizontal (for ground radar) or vertical (for space)—considering the purpose of ultimately simulating space-based $K_u/K_a$-band observations, we have used the horizontal geometry for the C-band and the vertical geometry for all the others.

The values of the parameter $K_w$ are based on the refractive index data for liquid water published in [12], except for W-band where we adopted the value used by the CloudSat team [13]. These values are summarized in Table I. It should be noted that discrepancies in the values $K_w$ only introduce errors of a constant factor (constant offset on the dBZ scale) and can be easily accounted for if necessary. The dielectric properties of snow are modeled as a mixture of ice and air (melting snowflakes are not considered in the scope of this paper). We used the refractive indices $m_i$ for ice by Warren and Brandt [14]. The refractive index for snow $m_s$ is obtained from those using Maxwell–Garnett mixing, which gives

$$\frac{m_s^2 - 1}{m_s^2 + 2} = \rho_s \frac{m_i^2 - 1}{m_i^2 + 2}$$

where $\rho_s$ and $\rho_i$ are the densities of snow and pure ice, respectively. The Maxwell–Garnett formula, traditionally formulated for sparse media, has been shown to be also accurate for dense materials in many cases [15].

Finally, for the particle size distribution, we use the standard normalized gamma model [16]

$$N(D) = N_w f(\mu) \left( \frac{D}{D_0} \right)^\mu \exp \left( -\left( 3.67 + \mu \right) \frac{D}{D_0} \right) \quad (3)$$

$$f(\mu) = \frac{6}{3.67^4} \frac{(3.67 + \mu)^{\mu+4}}{\Gamma(\mu + 4)} \quad (4)$$

with shape parameter $\mu$, intercept parameter $N_w$, and median volume diameter $D_0$. Applying this distribution, the reflectivity is computed from (1), using the radar cross sections from the T-matrix code, with numerical integration.

Although the snow model is only strictly valid for the specific assumptions given in Table I, it can be applied somewhat more generally. For different assumptions of the water dielectric factor $K_w$, simple rescaling of data is sufficient to apply the model, and assigning slightly different exact frequencies to the frequency bands (for example, using 95 GHz instead of 94 GHz for W-band) is unlikely to cause any major discrepancy in the results. As for different geometries, the C-band results are not very sensitive to a change in viewing angle due to the small size of the scatterers compared with the wavelength. For shorter wavelengths, varying the viewing geometry can cause a change in the correct mapping functions, but it should be noted that, in addition to spaceborne radars, the vertical geometry also applies to vertically pointing airborne radars and upward-looking ground radars.

B. Cross-Frequency Simulations

We have used the simulation described earlier with properties observed in real snowfall to examine the relationship of reflectivity values at different bands. The approach is similar to the cross-frequency technique used in [17] as well as the hydrometeor size retrievals in, e.g., [18]–[23]. The results, shown in Fig. 1, are generated by sampling a range of microphysical and drop-size distribution parameters. The parameter ranges were as follows:

$$0.4 < r < 0.8 \quad (5)$$

$$0^\circ < \sigma_\beta < 40^\circ \quad (6)$$

$$0.05 \text{ mm} < D_0 < 8.0 \text{ mm} \quad (7)$$

$$10^2 \text{ mm}^{-1} \cdot \text{m}^{-3} < N_w < 10^6 \text{ mm}^{-1} \cdot \text{m}^{-3} \quad (8)$$

$$-1 < \mu < 4. \quad (9)$$
Fig. 1. Cross-frequency relations from C- and W-bands to $K_u$- and $K_a$-bands. The shapes of markers show the mass-dimensional relation used: Squares for (10), diamonds for (11), and circles for (12). The shaded area gives the spread of the relation, and the dashed line gives the same relation with the assumption of Rayleigh scattering. (a) $C \rightarrow K_u$. (b) $C \rightarrow K_a$. (c) $W \rightarrow K_u$. (d) $W \rightarrow K_a$.

The individual values from the parameters were generated from the distributions, assuming a uniform distribution except for $N_w$, for which $\log_{10} N_w$ was uniformly sampled. $D_{\text{max}}$ was selected to be 3.0$D_0$.

The ranges of the axis ratio $r$ and the canting-angle standard deviation $\sigma_\beta$ were based on the values used in recent studies [24]–[27]. The drop-size distribution parameters are also similar to (and, to cover the extremes better, somewhat expanded from) those found in theoretical and experimental research published for snowfall [21], [28], [29]. For the selection of the shape parameter $\mu$, little quantitative reference material is available, but the experiments in the previously referenced studies indicate that the exponential distribution ($\mu = 0$) describes most snowfall with good accuracy. Where applicable, the parameter ranges are also consistent with those used for rainfall in [17].

For snow density, we used three different mass-dimensional relationships that have been published for dry snowflakes. These relationships exhibit considerable variability, and several of them are compared in [30]; we use this comparison to select mass–diameter relations at different ends of the range. From those, the relation from [31]

$$m = 0.0022D_m^{2.0}$$

(10)

was chosen to represent light snow. $D_m = D/\sqrt{r}$ is the maximum diameter of the snowflake and is given in centimeters in all size–mass relationships presented here; the mass in given in grams. Considerably higher densities are given by the relationship interpreted in [30] from the results of [32]. Using the assumed relation $D_s^3 = 0.15D_m^3$ of the mean diameter $D_s$ of snowflakes to the maximum diameter from [30], we get (assuming $r = 0.6$ for the purposes of deriving the formula)

$$m = 0.0027D_m^{1.5}.$$  

(11)
Additionally, a more recent piecewise defined formula

$$m = \begin{cases} 
0.003D_m^{2.0}, & D_m \leq 0.2 \text{ cm} \\
0.006TD_m^{2.5}, & 0.2 \text{ cm} < D_m \leq 2.0 \text{ cm} \\
0.004TD_m^{3.0}, & D_m > 2.0 \text{ cm} 
\end{cases}$$

(12)

compiled from earlier results in [26] was chosen to represent the best current knowledge.

We decided not to use prior climatological information other than the rough ranges of parameters, given by (5)–(9), in the sampling distribution. A more exact approach would probably result from drawing points from global statistical distributions of snow parameters, but data that could be used to derive such distributions are scarcely available and dependent on the climate in which they were obtained. Additionally, testing with combinations of values that are beyond the limits of the range for strictly realistic snow enables us to examine the sensitivity of the model to such statistical outliers. The only direct statistical information that we used was data from several years of radar measurements in Finland [33], [34], which shows that the reflectivity of snowfall at the C-band is limited to $-25 \text{ dBZ} < Z_C < 35 \text{ dBZ}$. Samples where the C-band reflectivity was outside this range were discarded.

Fig. 1(a)–(d) shows a large difference between the different mappings in the scattering of the values around the average (marked with the solid lines). It is made evident by Fig. 1(a) that the $K_a$-band reflectivity can be quite reliably estimated from the C-band throughout the range of snow parameters. This is not surprising, as the $K_a$-band wavelength of 22 mm guarantees that the scattering process is mostly within the Rayleigh scattering range, where the reflectivities are equal, because radar reflectivity is defined such that, for particles that are much smaller than the wavelength, it gives equal values independent of the particle size or the wavelength.

Predicting $K_a$-band observations is somewhat more complicated. The reflectivity can be estimated from the C-band up to 10–15 dBZ [Fig. 1(b)], but the scatter plot diverges rapidly beyond that. Fortunately, the deviation of the points from the mean for the W-band to $K_a$-band comparison [Fig. 1(d)] starts to converge at values above about 15 dBZ. Thus, W-band rather than C-band data could be used at high values of reflectivity to predict $K_a$-band measurements. A more accurate prediction can be made by combining the two values into a 2-D lookup space. Fig. 2 shows that, with knowledge of C-band reflectivity and the DFR, defined as the ratio of the reflectivities at the two bands, the reflectivity at the $K_a$-band can be solved practically unambiguously. Using the DFR is not strictly necessary, and one would obtain the same results by using the W-band reflectivity on the vertical axis in Fig. 2; we have used the DFR here for reasons of visual clarity and simplicity of an approximating function (Section II-C). The same concept can be used for other frequency combinations, such as using the $K_a$-band as the lower frequency instead of the C-band (see Section IV).

Dual-frequency data can, similarly to the technique in [35], also be used to estimate the attenuation of the signal by snowflakes at the high-frequency bands by using the reflectivity at a lower unattenuated frequency and the DFR of the two given bands. Although lower than in rain or the melting layer, the attenuation can be significant in heavy snowfall at the W-band and, to a lesser extent, at the $K_a$-band. With the parameters used in this study, attenuation at W-band was found to have a maximum of approximately 1 dB/km. The attenuation reaches nonnegligible values ($A > 0.1 \text{ dB/km}$) already at relatively modest reflectivities of 10–20 dBZ. At the $K_a$-band, the attenuation was at most 0.2 dB/km and usually much smaller. For C- and $K_a$-bands, the attenuation by snowflakes is insignificant and has been neglected in this paper. We note that attenuation may also be caused by other sources on the path of the radar.
beam that are not considered by this method, such as water vapor and supercooled water, but these can be quantified by other means; for a nadir-looking radar, the main source is cloud water attenuation which can be up to 1–2 dB.

We also considered the effects of multiple scattering in snowfall. It affects radar measurements when scattered radiation returns to the radar after scattering several times and thus traveling a longer distance than single scattered radiation, which causes an error in radar ranging. It is typically present together with heavy attenuation and can be thought of as a return of a part of the attenuated radiation. Simulations of extreme snowfall cases with a freely available multiple scattering code [36] indicated that, even in the presence of heavy attenuation, multiple scattering can be neglected for dry snowflakes at all of the frequencies used in this paper.

C. Empirical Fits

In this paper, the relations derived earlier have been applied to data (see Section III-B) using lookup tables corresponding to the averages shown in Fig. 1. For ease of future application, we have also derived approximate mapping functions for the various relations.

The frequency-to-frequency maps (Fig. 1) have been estimated as polynomial fits of the fourth degree. Transforming from reflectivity $Z_A$ (dBZ) to $Z_B$ (dBZ), where $A$ and $B$ correspond to the bands used, the map is given by the function

$$Z_B = Z_A - d_{AB}^R + f_0 + f_1 Z_A + f_2 Z_A^2 + f_3 Z_A^3 + f_4 Z_A^4.$$  \hspace{1cm} (13)

The coefficients $f_0 \ldots f_4$ for each relation are in Table II(a). With $f_0 = \ldots = f_4 = 0$, the fit describes the relation at the Rayleigh limit. The constant $d_{AB}^R$ is equal to the DFR (in the decibel scale) in the reflectivities in the Rayleigh regime and is due to the variation of the refractive indices of ice and water at the different bands. The value of this constant can be derived from the definitions of reflectivity as

$$d_{AB}^R = 10 \log_{10} \left( \frac{|K_{BI}^B|^2 |K_{AI}^A|^2}{|K_{BI}^I|^2 |K_{AI}^I|^2} \right)$$  \hspace{1cm} (14)

where the superscripts $A$ and $B$ denote the values for the corresponding bands and $K_i = (m_i^2 - 1)/(m_i^2 + 2)$ is the dielectric factor for ice. Including the terms $Z_A - d_{AB}^R$ in the fit allows the use of the empirical function for different values of $K_w$ simply by changing the value of $d_{AB}^R$. In Table II(e), we also give the values of $d_{AB}^R$ used to generate the plots.

The spread of one standard deviation $\sigma$, in decibel units, of the reflectivity [the shaded area in Fig. 1(a) and (b)] around the value can be approximated by

$$\sigma(Z_B) = \frac{a}{1 + ((Z_A - Z_0)/b)^2}$$  \hspace{1cm} (15)

using the values of the constants $a$, $b$, and $Z_0$ from Table II(b) ($Z_A$ and $Z_0$ are given in logarithmic dBZ units).
(a) Coefficients for the frequency-band relations \([13]\), valid for \(-25 \text{ dBZ} < Z_A < 35 \text{ dBZ}\) (b) Spread of the frequency-band relations \([15]\), valid for \(-25 \text{ dBZ} < Z_A < 35 \text{ dBZ}\) (c) Combined frequency maps \([16]\) (d) Reflectivity-attenuation maps \(Z_A, Z_B \rightarrow A_B\) \([17]\) (e) Values of \(d_{AB}_R\) \([14]\) used in this paper

For the dual-band to single-band relation, we propose the following function:

\[
Z_B = Z_A - d_{AC} R + a \left( \text{DFR}_{AC} - d_{AC} R \right) + b \left( \text{DFR}_{AC} - d_{AC} R \right)^\alpha
\]

(16)

where the constants \(b, c,\) and \(\alpha\) and the fit error are given in Table II(c) and the reflectivity and DFR are, as before, in dBZ and dB units, respectively. The value of this function is usually within 1 dBZ from the true value; an error analysis for predicting the \(K_a\)-band reflectivity from C/W-bands is shown in Fig. 3.

The specific attenuation at band B can be computed from reflectivity at bands A and B with a similar fit

\[
\log_{10} A_B = a + b \left( Z_A + 10 \log_{10} R^{AC} \right) + c \left( \text{DFR}_{AC} - d_{AC} R \right)
\]

\[
+ d \left( \text{DFR}_{AC} - d_{AC} R \right)^{1/2}
\]

(17)

where the logarithm of \(A_B\) is taken in with respect to 1 dB/km and the reflectivity and DFR values are again in logarithmic units. The accuracy of this estimation is comparable to (16). The constants \(a-d\) are given in Table II(d).

III. APPLICATION ON COMBINED GROUND AND SPACE MEASUREMENTS

A. Measurement Setup

The research radar of the University of Helsinki (referred to as “Kumpula radar” in this paper) is a 5.6-GHz (C-band) dual-polarized Doppler research radar that has been in operation since December 2004. It is located on the roof of the Physicum building in Kumpula, Helsinki, at \(60^\circ 12'26''N\) \(24^\circ 57'78''E\), 59 m above the mean sea level and 30 m above the ground. During the year 2009, the Kumpula radar has been carrying out measurements synchronized with CloudSat satellite overpasses.

CloudSat uses a nadir-looking 94-GHz (W-band) radar \([13]\) that can measure both clouds and precipitation. Because of its high sensitivity, CloudSat has proved to be a reliable tool for...
space-based measurements of light rain and snowfall, which is also less susceptible than heavy precipitation to the problems posed at high frequency by attenuation and multiple scattering effects.

CloudSat orbits in the Afternoon Constellation (“A-train”) of Earth observation satellites at an inclination of 98.2° [37]. Due to the high latitude (60° N), the ground track of the near-polar orbiting CloudSat passes close to Helsinki relatively frequently, twice per eight days on four repeating paths at an approximately 30- or 50-km distance at closest approach to the radar (see Fig. 4 for an illustration). Since precipitation is not always present, this has produced about one usable profile per month on the average, depending on the weather and the uptime of radars.

To provide simultaneous measurements, the ground-based radar is preprogrammed using CloudSat orbit predictions [38]. The radar then performs a sector scan in the direction of the satellite ground track, coinciding as closely as possible with the overpass. This is implemented as a radar sweep covering a sector of 100° azimuth angle centered on the point of closest satellite approach. The entire scan takes roughly 4 min and is started 2 min prior to the closest approach. The scans are performed using 15 elevation angles from 0° to 20°, with a PRF of 600 Hz and 32 sample integrations with a 0.5° sample width. A cross section is produced from the 3-D radar data to match spatially the 2-D data provided by the CloudSat nadir-looking radar.

The main limitations of the measurement setup are related to the geometry between the satellite, the ground-based radar, and the observed cross section. First, the precipitation cannot be measured near the surface, due to the curvature of the Earth for the ground-based radar and the ground echo for CloudSat. In our case, the lower limit of the radar beam height is about 0.35 km (for the closer overpasses) or 0.7 km (for the farther ones), leaving the CloudSat ground echo top of approximately 1 km as the limiting factor.

Second, a degree of cross-section mismatch is inevitable with the equipment used. Since performing the sector scan with the Kumpula radar takes minutes, rather than seconds needed by CloudSat to cover the same area, there is a slight time difference, varying with location, between the measurements. This introduces spatial misalignment due to the movement and change in the shape of the observed weather pattern. For a typical wind speed of 10 m/s and a time difference of 1.5 min, the drift distance is 900 m, smaller than the CloudSat along-track resolution of roughly 1.7 km [13]. For a more extreme case of 35-m/s wind and 3-min difference, the displacement can be up to 6 km, which can cause mismatch-related errors depending on the horizontal homogeneity of the precipitation structure. Together with the difference of the resolutions of the ground radar and the satellite, the mismatch gives rise to some artifacts, primarily at the edges of features. We estimated the error by producing alternate cross sections, shifted from the original according to the mean wind in the area, as given by atmospheric sounding data, and the maximum time difference, providing a worst-case estimate. The typical errors were quite small (rms of 1.0–1.8 dB, median of 0.5–1.0 dB) even for strong winds (up to 35 m/s at high altitudes). This is as expected because snowfall is known to be quite spatially homogeneous compared with rain [39]. It was found that the wind drift effect can cause occasional large (up to 10 dB) errors at the feature edges, although these can be considered as statistical outliers and are easily spotted by a human observer. Nonetheless, filtering or interpolation techniques are likely needed to eliminate these artifacts if large-scale automatic processing of the dual-source data is performed.

B. Snowfall Case Studies

The frequency mapping methods introduced in the previous sections were used to study overpasses of the CloudSat satellite, simultaneously observed by the Kumpula radar. We examined one case in particular: snowfall measurements from April 4, 2009, shown in Fig. 5.

The CloudSat overpass took place at 11:07 UTC. The Kumpula radar had started the sector volume scan centered on the closest CloudSat approach at 11:06 UTC. At the time of observations, a widespread rainfall with a shallow freezing level took place, Helsinki Testbed surface temperature measurements showing temperatures ranging from 3 °C to 5 °C west of the radar and 1 °C to 2.5 °C east of the radar. The closest CloudSat overpass point was 35 km at an azimuth of 70° from the Kumpula radar. Given the surface temperature observations and the 12 UTC sounding, we can conclude that a melting layer was present just below the minimum observable height of CloudSat. The sounding was carried out about 100 km northwest of the radar and showed the melting level height at 750 m above mean sea level. It can be seen that Kumpula observations do not exhibit a typical bright band type of signature even at lower altitudes. This was also verified by an analysis of dual-polarization radar observations. Therefore, in our analysis, we treat these observations as a snowfall case.

Fig. 6 shows the structure of the correlation of the two measurements. We can identify some significant features: The
Fig. 5. Application of the mapping methods on combined ground and space data. The case is from April 4, 2009.

Fig. 6. Scatterplot of the simulated C- and W-band radar reflectivities [circle, square, and diamond markers as with Fig. 1(a)-(d)] and the values measured in the April 4, 2009 snowfall event (plus signs). Only the points left after removing the data considered invalid (i.e., the same as those shown in Fig. 5) are shown. The approximated contours give the diameter \( D_0 \) of the particle distribution; the edge of the shaded area is the small particle (Rayleigh) limit. The points in the shaded area cannot be interpreted physically and must instead be attributed to noise, partial beam filling, or other errors.

observations are well correlated but scattered; the spread of the points is centered approximately 2 dB below the Rayleigh scattering line but well above the mean simulated line; at the low end of the reflectivity range, the spread is larger; and at the high reflectivities, some large mismatches are found. The great majority of the points are contained within the area predicted by the simulation parameters, and the concentration of the points in the small-particle region indicates that the snowflakes in the event were small in size, consistent with the low radar reflectivity. The larger spread of points at low reflectivity can be explained by higher sensitivity to noise in that range and the outliers at high reflectivity by clutter or feature edge mismatches (as explained in Section III-A).

At the low reflectivity values of this case, W-band attenuation by the snowflakes can be neglected, but attenuation due to cloud liquid water and water vapor may still be present. Using the low-reflectivity end of the scatterplot, we can attempt a rough quantification of this attenuation. As the true reflectivity values at the low reflectivity limit are expected to be given by the Rayleigh scattering assumption for both frequencies, the difference of these values may be used to infer the amount of attenuation present. Although the values are highly scattered at the low end due to increased susceptibility to noise, the mean does approach very close to the Rayleigh limit, indicating that a large amount of attenuation is not present. Accordingly, data in Figs. 5 and 6 were not adjusted to compensate for attenuation.

Synthetic \( K_u \)- and \( K_a \)-band radar observations were produced for the event from the analysis of the C- and W-band measurements. The \( K_u \)-band radar observations can be directly derived from the C-band measurements using (13), as was shown in Section II-B. Therefore, the synthetic \( K_u \)-band measurements exhibit similar features and are thresholded in the same way as the Kumpula radar cross sections. As expected, there is not a large difference between the C-band measurements and the simulated \( K_u \)-band values, considering that the cases represent light snowfall with low reflectivity, and according to Fig. 1(a), the C-band values are nearly equal to those at the simulated bands.

\( K_a \)-band radar observations, on the other hand, were derived from a joint analysis of Kumpula radar and CloudSat observations with (16). As can be seen, the W-band data appear to be quite different from the weather radar observations and the simulated observations. This difference is due to a change in scattering processes: In this case, there is a transition from Rayleigh to Mie scattering between the \( K_a \)-band and the W-band, and this reduces the observed reflectivity at the W-band significantly. As can be seen from Fig. 1(b), \( K_a \)-band
radar observations almost perfectly match the C-band observations for reflectivity values smaller than 10 dBZ. Therefore, it is not surprising that the synthetic $K_u$-band observations do not differ much from the C-band measurements. Furthermore, the differences only occur in high reflectivity areas, where the weather radar reflectivity exceeds roughly 10 dBZ. For the $K_u$-band, the resulting reflectivity field is limited by both the CloudSat and Kumpula detection thresholds, since both CloudSat and Kumpula radar observations were used to generate these data.

IV. APPLICATION ON DATA FROM THE WAKASA BAY EXPERIMENT

In January and February 2003, NASA conducted a precipitation measurement field campaign in the Wakasa Bay area of the Sea of Japan [40]. In the 12 measurement flights of this campaign, a NASA P-3 aircraft carried, among other instruments, two radars: the Airborne Second Generation Precipitation Radar (APR-2) with matched beams at the $K_u$- and $K_a$-bands [41] and the Airborne Cloud Radar (ACR) at the W-band [42].

APR-2 and ACR are somewhat different instruments. APR-2 is a scanning radar with a beamwidth of 3.8° for the $K_u$-band and 4.8° for the $K_a$-band. Each scan starts 25° left of the nadir and ends 25° right of the nadir, consisting of 24 rays with a range bin spacing of 30 m. In contrast, ACR points statically at the nadir with a beamwidth of 0.8° and a typical vertical resolution of 120 m. Thus, a data matching, similar to that described in the previous section, needs to be performed. Due to the different coordinate specifications used by the radars, it was found simplest to use the distance from the aircraft (rather than the surface) as the common vertical coordinate. As the horizontal coordinate, we simply used the measurement time. The ACR data, having the lower vertical resolution, were mapped to the APR-2 coordinates using linear interpolation. From the APR-2 data, results from beam #12, pointing downwards in the aircraft reference frame, were used. We verified the matching of the data visually by comparing the positions of features such as the melting layer brightband and the ground echo in $K_u$/$K_a$-band cross sections and the interpolated W-band data.

We used the data from the experiment to validate the $K_u$/$W$-band to $K_a$-band mapping relation given in (16). As a test case, we used the measurements from January 29, 2003. On that day, widespread snowfall occurred over land and ocean [41], providing a clear snowfall scenario without distracting features such as the melting layer and the underlying liquid rain. The radar reflectivity patterns were characterized by clear features in both horizontal and vertical directions, facilitating the data fitting procedure. Data were used down to the minimum altitude where ground echoes were clearly absent, roughly 0.5 km. We did not attempt to remove cloud echoes, due to both the difficulty of identifying clouds from the data and the intent to test the sensitivity of the model to clouds.

In Fig. 7, we show an example of applying the model on the data. The ACR measurements were first interpolated to the coordinates of the APR-2 $K_u$-band data [Fig. 7(a)]. The attenuation correction for the W-band was then computed using a simple top-to-bottom algorithm. While numerically unstable, this straightforward method was used for its simplicity and did not appear to cause any artifacts due to instability. The DFR of $K_u$- and W-bands was computed for both the corrected [Fig. 7(b)] and uncorrected W-band cross sections, and the estimated $K_a$-band reflectivity [difference of the estimate and the measurements given in Fig. 7(c) and (d)] was computed from both of these using (16). The observed $K_a$-band reflectivity was also subjected to the attenuation correction described previously before comparing with the estimated values; this implies that Fig. 7(d) compares the true (unattenuated) reflectivities.

The method can be seen to predict the $K_a$-band reflectivity with fairly good accuracy. If attenuation correction was not performed, a systematic bias (defined as the mean difference of the logarithmic values) of 0.66 dB was found for the prediction, with a median absolute error of 0.93 dB. Using the attenuation correction, these values increased to 0.92 and 1.03 dB, respectively. However, the elimination of the gross underestimates at the bottom left of Fig. 7(c) by the attenuation correction clearly shows that the actual quality of the prediction is greatly improved by the correction—the increase in bias is due to the fact that the previously underestimated areas no longer compensate for the positive bias found elsewhere. The bias is within the spread of the scatter shown in Fig. 3 and can thus be explained by the deviation of the microphysical properties of snow from the average, although other factors such as calibration differences between the radars may also contribute to it.
V. Discussion and Summary

Dual-band radar data can be used to simulate precipitation radar observations of snowfall at different bands accurately. By using scattering simulations at typical parameters of natural falling snow, one can derive frequency maps of reflectivity and use these maps on measured data to predict radar measurements at frequencies where no direct observations are available. A dual-frequency attenuation estimation scheme can improve the reliability of the approach.

Motivated by the use $K_u/K_u$-band dual-frequency radar in the GPM satellite, we have demonstrated the effectiveness of this approach for the case of approximating observations at these bands from C-band data from a ground-based radar and W-band spaceborne measurements from the CloudSat satellite. The targets of these observations were matched by scanning along the CloudSat ground track with the ground-based radar during the satellite overpass. In another test case, we validated the estimation of $K_u$-band radar data from the $K_u$-band and the W-band using airborne radar data from the Wakasa Bay field campaign, with concurrent $K_u$-band measurements used as a reference.

In this paper, we have also provided empirical functions for the cross-frequency maps, which can be used to predict values of radar reflectivity and attenuation at frequencies where no measurements are available and to estimate radar attenuation using dual-frequency measurements. This approach can be used for the planning of future missions and for validation of current and future radars in the absence of existing data on the given frequency bands.

Recently, concerns have been raised regarding the effect of the complex nonspherical shapes of real snowflakes on radar scattering [43], [44]. These studies have shown that the backscatter cross sections of individual snowflakes can differ by orders of magnitude from spherical or spheroidal approximations. The applications of the spheroidal model on experimental measurements considered here, particularly the Wakasa Bay data, do not seem to support such extreme differences. This implies that the simplified model can still be usable when interpreted properly. An important distinction between the aforementioned studies and the present one is that we consider snowflakes only as large ensembles of different sizes and shapes. In this context, the spheroid model appears to be able to function as a good approximation of the average over the ensemble. Identifying how such “effective soft spheroids” can be defined for a given particle distribution is certainly an interesting topic for further study, with implications for how the particle diameter (and, inevitably, the density) is interpreted in this context. The method described in this paper can be used to test these models and detect deviations from the spheroidal approximation.

Aside from the issue with particle shapes, there are four significant sources of error that determine the limitations of the approach. The first type is related to the effect of the assumed global distributions of the physical parameters of snow particles such as size, shape, and constitution. We have used only roughly realistic ranges, rather than proper $a$ priori snow parameter distributions (which are climate specific and presently not readily available), and the error in these assumptions can propagate to the scatter of points and further to the results. According to a sensitivity analysis we conducted, the resulting mapping functions are not affected more than slightly by realistic changes to the parameter ranges. The second source is the uncertainty that is induced to the final result by the spread of the scatterplot around the mean; however, this is greatly alleviated by the use of radar reflectivities at two frequencies simultaneously. With both data available, the $K_u/K_u$-band observations can be simulated with a typical accuracy of 1–2 dB. Third, for combined ground and space measurements, the limited scanning speed of the ground-based radar compared with the satellite which flies over the area in seconds, as well as the incomplete overlap of the ground and space radar bins, cause the spatial and temporal matching of the two data to be imperfect, which typically causes errors which are as big as or smaller than those from the previous two sources, but due to mismatches at the edges of features, can cause occasional large errors in the form of artifacts that are quite conspicuous to human observers. In the case of airborne measurements, this potential for error is smaller as the radars are colocated in the aircraft, but potential errors from bin mismatch still apply for radars without matched beams. Finally, the uncertainty of the original radar measurements, with a typical magnitude of roughly 1 dB, is itself inevitably reflected in the simulated results as well. As the slope of the mapping functions is close to 1, this error is not amplified greatly by the mapping, and the error due to this source is at least not larger than any of the aforementioned error contributors and is likely to be comparable with the uncertainty of the target radar. Overall, it can be said that the method performs well for the great majority of the time, but can be prone to unexpected errors that require the experience and heuristic skill of the observer to recognize.

An important potential future improvement to the method is the development of techniques to match radar cross sections from temporally and spatially separated sources in a robust way, particularly for the purpose of automated data processing. Advancements in the understanding of global and regional climatological statistics of snowfall, which is currently incomplete, are also expected to contribute to the accuracy of determining radar observables at given frequency bands from data at different bands.

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REFERENCES


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