

RESEARCH ARTICLE

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Key Points:

- CloudSat blind zone effects were investigated using Micro Rain Radar profiles
- Blind zone can lead to underestimation of reflectivity and precipitation amount
- Reducing the blind zone by 50% does not improve results in all aspects

Supporting Information:

- Readme
- Figures S1–S3

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How does the spaceborne radar blind zone affect derived surface snowfall statistics in polar regions?

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Abstract Global statistics of snowfall are currently only available from the CloudSat satellite. But CloudSat cannot provide observations of clouds and precipitation within the so-called blind zone, which is caused by ground-clutter contamination of the CloudSat radar and covers the last 1200 m above land/ice surface. In this study, the impact of the blind zone of CloudSat on derived snowfall statistics in polar regions is investigated by analyzing three 12 month data sets recorded by ground-based Micro Rain Radar (MRR) at the Belgian Princess Elisabeth station in East Antarctica and at Ny-Ålesund and Longyearbyen in Svalbard, Norway. MRR radar reflectivity profiles are investigated in respect to vertical variability in the frequency distribution, changes in the number of observed snow events, and impacts on total precipitation. Results show that the blind zone leads to reflectivity being underestimated by up to 1 dB, the number of events being altered by $\pm 5\%$ and the precipitation amount being underestimated by 9 to 11 percentage points. Besides investigating a blind zone of 1200 m, the impacts of a reduced blind zone of 600 m are also analyzed. This analysis will help in assessing future missions with a smaller blind zone. The reduced blind zone leads to improved representation of mean reflectivity but does not improve the bias in event numbers and precipitation amount.

1. Introduction

As an integral part of the water cycle in polar regions, snowfall is extremely difficult to capture at the relevant spatial scale and with sufficient accuracy [Levizzani *et al.*, 2011]. Surface observations by in situ sensors are sparse and available only over land [Schneider *et al.*, 2014], and snowfall measurements in polar regions are particularly affected by wind-induced errors such as undercatch or overcatch from blowing snow [Yang *et al.*, 1999; Knuth *et al.*, 2010].

An important source of spatially extensive measurements of snowfall are remote sensing observations from space, such as microwave radiometer [Levizzani *et al.*, 2011] or the radar of the CloudSat satellite [Stephens *et al.*, 2008]. CloudSat allowed snowfall climatologies for polar regions up to 82°N/S to be derived for the first time [Liu, 2008a; Kulie and Bennartz, 2009; Palermé *et al.*, 2014]. However, accurately determining the snowfall rate (S) from the observed profile of the equivalent radar reflectivity factor (Z_e) is challenging due to the microphysical and microwave scattering uncertainties in the conversion of Z_e to S [e.g., Hiley *et al.*, 2011] and the inability to reliably measure Z_e close to the surface when using satellite radar systems. Our research addresses this latter point by studying how this “blind zone” affects observations and what impact a reduced blind zone would have.

The vertical extent of this blind zone is largely determined by the surface type and is smaller over the ocean and larger over land or sea ice [Durdén *et al.*, 2011a]. For CloudSat, it has been shown that over land, the received signal is free from ground clutter only from the fifth range bin above ground level (1200 m agl, hereafter $H_{CS} = H_{\text{CloudSat}}$) [Marchand *et al.*, 2008]. New spaceborne radars such as the recently launched dual-frequency radar on board the Global Precipitation Measurement (GPM) Core Observatory [Hou *et al.*, 2014] or the Cloud Profiling Radar (CPR) on board the upcoming Earth Clouds, Aerosols, and Radiation Explorer (EarthCARE) mission [Gelsthorpe *et al.*, 2010; Donovan *et al.*, 2013] are expected to achieve smaller blind zones ranging between 600 m and 1000 m above surface (depending on radar operation mode). Other proposed missions such as the Polar Precipitation Measurement Mission [Joe *et al.*, 2010] or the Aerosol/Cloud/Ecosystems (ACE) mission [Durdén *et al.*, 2011b] are heading to blind zones in the order of 100 to 200 m.

These observational blind zones are likely to introduce errors in the derived statistics of snowfall frequency and S for surface level if the snowfall properties are significantly altered within the blind zone. For example, snowfall might be underestimated or completely missed, if the snow cloud is shallow. The opposite extreme of snowfall overestimation would be snowfall that is detected at the top of the blind zone but completely sublimates on its way down to the surface. Besides these extreme scenarios, the snowfall properties observed at the top of the blind zone might also be altered within the blind zone due to microphysical processes or wind shear. For midlatitude systems, a general increase of the radar reflectivity factor toward the ground in the range of 3 to 7 dB km⁻¹ was found and has been associated to aggregation and depositional growth [Fabry and Zawadzki, 1995; Liu, 2008a; Matrosov and Battaglia, 2009; Wolfe and Snider, 2012]. While there is large variability, reflectivity gradients are found to generally increase with higher temperatures. Henson *et al.* [2011] and Stewart *et al.* [2004] studied ground-based radar data for the Canadian Arctic region and found radar reflectivity in the lowest 2 km above ground to be nearly constant. However, some cases revealed increasing as well as decreasing radar reflectivities in the range of 5 dB km⁻¹ toward ground depending on the thermodynamic structure of the lowest layers and the intensity of snowfall. Despite its blind zone limitations, CloudSat still detects significantly more light snowfall than the ground-based precipitation radar networks do because of its higher sensitivity [Smalley *et al.*, 2014].

The goal of this paper is to investigate the impact of the blind zone on snowfall statistics at three polar sites and to estimate whether statistics derived from observations above the blind zone are biased in comparison to statistics taken at the surface. One way of investigating the impact of the blind zone is to directly compare CloudSat overpasses over a site where the lower reflectivity profile is measured with results from a ground-based cloud radar. However, these direct comparisons can be strongly affected by the difference in observed radar volumes or the horizontal displacement of the satellite overpass from the location of the ground-based radar. These complications can be avoided by statistically comparing satellite and ground-based radar observations [Protat *et al.*, 2009, 2010]. Alternatively, variations of radar reflectivity within the blind zone using only ground-based observations can be investigated. We use the observations at the top of the blind zone from the ground-based radars as a reference for the statistics that a theoretically perfect overpassing and volume-matched satellite radar would provide. In this way, we avoid any approximation due to temporal, spatial, or radar volume mismatch while the generally expected changes (e.g., underestimation/overestimation of snowfall) within the blind zone are assumed to appear in both the real satellite observations and the ground-based measurements.

In this study, we apply this method to long-term radar observations from observational sites in Antarctica and Svalbard. These ground-based observations were carried out using the Micro Rain Radar (MRR), a compact and lightweight 24 GHz frequency-modulated continuous wave (FMCW) radar [Klugmann *et al.*, 1996].

2. Study Area

We use column measurements from the Princess Elisabeth station (PE) in East Antarctica and from Ny-Ålesund (NÅ) and Longyearbyen (LY) in Svalbard, Norway, as exemplary data sets for polar regions. To the authors' knowledge, no other ground-based precipitation radar is currently operating in Antarctica and the data sets from NÅ and LY are the first of their kind in Svalbard. Because the setup and the results in LY were similar to NÅ and both sites are only 110 km apart, results for LY are only presented in the supporting information. Nevertheless, comparison of both sites is important when assessing how representative the study is.

2.1. Princess Elisabeth Station, East Antarctica

The only current precipitation radar over the Antarctic ice sheet is installed as part of the meteorological-cloud-precipitation observatory that has been operating at the PE station since February 2010 [Gorodetskaya *et al.*, 2014a] (<http://ees.kuleuven.be/hydrant>). The station is built on Utsteinen Ridge located north of the Sør Rondane Mountains in the eastern part of Dronning Maud Land (DML) at the ascent to the East Antarctic plateau (71°57'S, 23°21'E, 1392 m above mean sea level (amsl), 173 km from the coast, Figure 1a). Together with other remote sensing instruments, the MRR precipitation radar is installed on the roof of the PE station about 10 m above the snow surface (Figure 2a). An automatic weather station providing meteorological data is located 300 m east of the PE base [Gorodetskaya *et al.*, 2013]. In this study, we use year-round measurements available for the full year 2012.

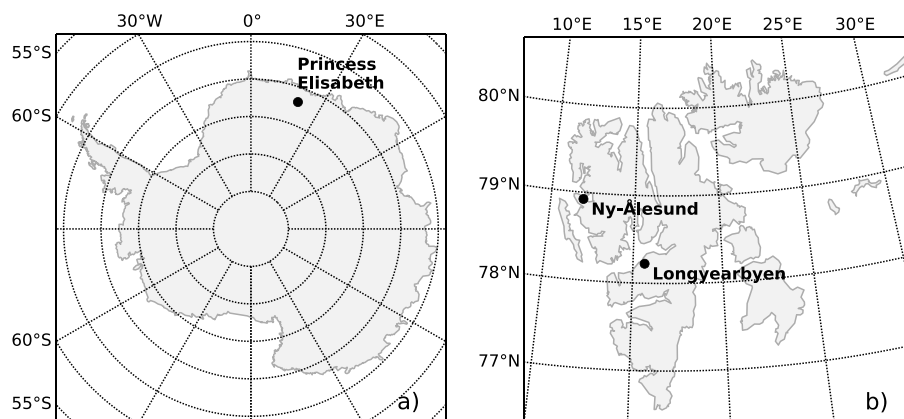


Figure 1. Map of the location of Princess Elisabeth station in (a) East Antarctica and (b) of Ny-Ålesund and Longyearbyen in Svalbard, Norway.

The PE site is characterized by a relatively mild climate, mainly due to the favorable location for warm air advection associated with local intense cyclonic activity and a lack of drainage of cold air from the high plateau due to shelter from the Sør Rondane Mountain [Gorodetskaya *et al.*, 2013]. Two main meteorological regimes govern the weather at PE—the cold katabatic regime and the warm/transitional synoptic regime [Gorodetskaya *et al.*, 2013]. The katabatic regime is characterized by low wind speeds of predominantly south-southeasterly direction, strong near-surface temperature inversions, low specific humidity, and low incoming longwave fluxes indicative of clear skies. The synoptic regime at PE is in turn associated with the Southern Ocean cyclones passing near DML and bringing heat and moisture advection into the Antarctic ice sheet [Noone *et al.*, 1999; Schlosser *et al.*, 2010]. Precipitation at PE (only as snowfall) mainly occurs during synoptic regimes. The largest snowfalls observed at PE have been associated with narrow bands of enhanced moisture amounts (atmospheric rivers) directed into DML and surrounding sectors [Gorodetskaya *et al.*, 2014b]. Such rare large snowfall events contribute significantly to the total yearly snow accumulation and can explain the high interannual variability of snow accumulation at the PE site (from 23 up to 230 mm water equivalent per year [Gorodetskaya *et al.*, 2014a, 2014b]) and entire DML [Boening *et al.*, 2012; Lenaerts *et al.*, 2012].

2.2. Ny-Ålesund, Svalbard

Measurements in the Arctic were taken in NÅ (78.92°N, 11.93°E, 8 m amsl) on Spitsbergen, the largest island of the archipelago of Svalbard, Norway (Figure 1b). The climate in Svalbard is strongly influenced by the West Spitsbergen Current, which flows from the North Atlantic along the west coast of Spitsbergen and provides the largest input of sensible heat into the Arctic Ocean [e.g., Gammelsrød and Rudels, 1983]. Through mixing, some of this heat reaches the fjords in the west of the island and leads to a much milder climate

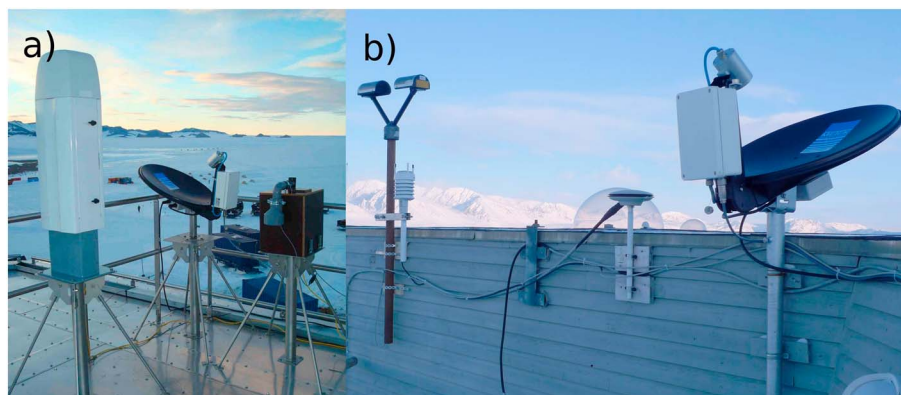


Figure 2. (a) Ceilometer (left), MRR (middle), and infrared pyrometer (right) on the roof of the Belgian Princess Elisabeth station in East Antarctica. (b) MRR (right) and Parsivel disdrometer on the roof of the Norwegian Polar Institute Sverdrup Station in Ny-Ålesund, Svalbard.

than in other locations at the same latitude (e.g., Greenland). Additionally, Svalbard lies in the North Atlantic storm track with general cyclogenesis and rapid cyclone deepening within the Icelandic Low and penetrating into the Arctic [Serreze *et al.*, 1997; Tsukernik *et al.*, 2007]. As a consequence, the atmospheric influence on the climate varies between moist and warm air masses coming from the Atlantic and dry and cold air masses coming from the Arctic. Precipitation in Svalbard can therefore be liquid as well as solid.

Ny-Ålesund lies at the shore of the Kongsfjord, which is 26 km long and up to 14 km wide. This position leads to a maritime climate, when the fjord is ice free, and higher precipitation, with an annual mean of 427 mm (for the period 1981–2010), than in the interior of Spitsbergen [Førland *et al.*, 2012]. The MRR was placed on the roof of the Sverdrup Research Station operated by the Norwegian Polar Institute (NP) in 8 m agl height (Figure 2b), 350 m away from the sea. Standard meteorological observations are taken from the weather station of the Norwegian weather service (World Meteorological Organization no. 01007, data available at <http://eklima.met.no>) located 100 m southeast of the NP station. Measurements were taken over a period of 1 year between 10 March 2010 and 15 March 2011.

2.3. Regime Classification

To investigate whether the impact of the blind zone on snowfall measurements depends on the type of snow event or ambient weather conditions such as stability or humidity, the data set is classified into different regimes. Sublimation of precipitation before reaching the surface (virga phenomenon) depends both on precipitation microphysical properties (such as particle size and terminal velocity) and on the ambient meteorological conditions [Clough and Franks, 1991; Wang *et al.*, 2004; Campbell and Shiobara, 2008; Evans *et al.*, 2011]. Dry meteorological conditions with a relatively warm subcloud layer will favor virga formation, whereas saturated and mixed subcloud layer will favor precipitation to the surface. The occurrence of either conditions was identified based on the classification of 3-hourly measurements of low-level temperature inversion (T_{inv}) and near-surface relative humidity with respect to ice (RH_i) in the vicinity of PE and NÅ stations.

Near-surface T_{inv} ($^{\circ}\text{C m}^{-1}$) is calculated as the difference between the air temperature and the skin surface temperature of snow. The air temperature is measured at the variable height (within 2–4 m) above the snow surface depending on snow accumulation. Surface temperature is calculated using measurements of outgoing and incoming longwave radiative fluxes (see Gorodetskaya *et al.* [2013] for details).

Prior to regime classification, the data were deseasonalized and standardized by subtracting monthly mean values and calculating z scores; the classification is based on hierarchical cluster analysis of RH_i and T_{inv} following Gorodetskaya *et al.* [2013]. Two main regimes were identified: the *dry stable regime* characterized by low RH_i and high T_{inv} , which represents 27% (41%) of all observations at PE (in NÅ), and for all other events the *wet unstable regime* with high RH_i and near-zero T_{inv} . MRR observations are grouped according to the two regimes. The regime classification was not applied to the data set in LY because no weather station was available in the immediate vicinity.

3. Radar Observations of Snowfall

In this study, MRR data are used to mimic CloudSat measurements (Table 1) while investigating how radar observations are altered in the blind zone.

3.1. CloudSat

CloudSat, which is part of the A-train satellite constellation, carries the Cloud Profiling Radar (CPR) operating at a frequency of 94.05 GHz (wavelength $\lambda = 3.2$ mm) [Stephens *et al.*, 2008; Tanelli *et al.*, 2008; L'Ecuyer and Jiang, 2010]. The raw measured return power is provided in the level 1 data product (1B-CPR) and converted into equivalent attenuated reflectivity factor, included in the 2B-GEOPROF data product [Tanelli *et al.*, 2008]. Near the surface, the reflectivity product is contaminated by ground clutter leading to significant power returns not related to hydrometeor occurrence [Marchand *et al.*, 2008; Tanelli *et al.*, 2008]. Although the latest 2B-GEOPROF version has a clutter reduction procedure, this procedure is less effective over land than over ocean and is particularly ineffective over mountainous terrain [Mace, 2006]. Consequently, snowfall rates at the surface derived from truncated 2B-GEOPROF profiles need to be approximated by snowfall rates at an elevated height of 1200 m (H_{CS}) above the surface, introducing an effective blind zone of 1200 m agl over land. To ensure equal processing of MRR and CloudSat (see section 3.3), Z_e data of the 2B-GEOPROF product are used even though a CloudSat snowfall product (2C-SNOW-PROFILE) [Wood, 2011; Wood *et al.*, 2013]

Table 1. Comparison of MRR and CloudSat's CPR

	MRR	CPR
Frequency (GHz)	24.23	94.05
Radar type	FMCW	Pulsed
Transmit power (W)	0.05	1820 (peak power)
Radar Power consumption (W)	25	–
Number of range gates	31	125
Range resolution (m)	100/60 ^a	485, resampled to 240
Measuring range (km)	0–3/0–1.8 ^a	0–25
Lowest usable range gate (m)	400/240 ^a	1200
Antenna diameter (m)	0.6	1.85
Beam width (two-way, 6 dB)	1.5°	0.12°
Minimum detectable Z_e (dBz)	–5 at 1200 m agl	–30
Integration time (s)	300	0.16
Cross-track resolution (km)	–	1.4
Along-track resolution (km)	–	1.8

^aDepending on configuration.

based on 2B-GEOPROF has recently been released. CloudSat is said to be overpassing a station if its nadir view is within a radius of 100 km. Based on this radius, we found 899 overpasses for the PE study area and 1579 overpasses for the NYA study area in the period between 2006 and 2013. Reflectivities of the lowest clutter-free range gate at H_{CS} as reported by the 2B-GEOPROF product are analyzed for each overpass.

3.2. Micro Rain Radar

The MRR (Micro Rain Radar) is a vertically pointing frequency-modulated continuous wave (FMCW) radar manufactured by METEK GmbH, Germany, [Klugmann *et al.*, 1996] operating at a frequency of 24 GHz ($\lambda = 1.24$ cm). Its low power consumption of only 25 W makes it particularly suitable for remote areas with limited power supply. The MRR was originally developed to measure rain [Peters *et al.*, 2002, 2005; Tridon *et al.*, 2011], but recent modifications in the MRR data processing [Maahn and Kollias, 2012] and comparisons to cloud radar observations [Kneifel *et al.*, 2011] have revealed that the MRR can also be used to study snowfall [Stark *et al.*, 2013; Gorodetskaya *et al.*, 2014a].

The MRRs were equipped with a 200 W antenna heating, but the heating was only used in NÅ when temperatures were around the melting point. At lower temperatures, the heating is unable to completely melt the snow, which causes the dish to glaciare resulting in a disturbance of the measurements. Presence of a thin layer of dry snow, instead, does not contaminate the measurements because attenuation by dry snow is very weak at K-band [Matrosov, 2007] and the snow is easily blown away by the ambient wind before a large layer can accumulate. The antenna dish of the MRR in NÅ was checked daily by the station staff to avoid glaciation of the dishes. The MRR at PE was regularly checked by the staff in summer and additionally supervised via webcam to check the status of the instrument during the unmanned period.

The MRR provides data at 31 range gates with a resolution of 100 m agl (60 m agl) at PE (in NÅ). The first two and the last range gate are extremely noisy and removed from the analysis. Because MRR's original software was designed for rain only, the alternative software package IMProToo designed for observations of snowfall by Maahn and Kollias [2012] has been used for this study. Due to the fixed MRR Doppler velocity range of 0 to 12 m/s, aliasing effects might occur at upward or very low fall velocities (as is common for snow). These effects can be corrected by IMProToo for all but the third MRR range bin (see Maahn and Kollias [2012] for details). Therefore, the third bin has also been excluded from the analysis so that an effective measurement range of 400 to 3000 m agl for PE (240 to 1800 m agl for NÅ) remains. Hereafter, the altitude of the lowest observation of the MRR will be called $H_{SF} = H_{SurFace}$. A contamination of the measurements due to blowing snow is unlikely at H_{SF} and above [Xiao *et al.*, 2000].

To increase MRR's sensitivity to around –10 dBz (–12 dBz) at PE (in NÅ), MRR data were averaged to 300 s. Even though this averaging reduces temporal and horizontal resolution (due to stronger advection effects), this reduction is justified as convective events with high temporal variability are rare at polar sites. Unfortunately, MRR's noise level varies significantly between different instruments and can vary by more than 5 dB in time, which leads to temporary lower sensitivities. To exclude the impact of a varying sensitivity on

Table 2. Equivalent Reflectivity-Snowfall Rate (Z_e - S) Relations From *Kulie and Bennartz* [2009] and Derived Conversion for Z_e From 35 GHz to 94 GHz

Particle Habit	Z_e (35 GHz)	Z_e (94 GHz)	Z_e Converted From 35 to 94 GHz
Three bullet rosette (LR3)	$24.04S^{1.51}$	$13.16S^{1.40}$	$0.69Z_e^{0.93}$
Aggregates (HA)	$313.29S^{1.85}$	$56.43S^{1.52}$	$0.50Z_e^{0.82}$
Snow (SS)	$19.66S^{1.74}$	$2.19S^{1.20}$	$0.28Z_e^{0.69}$

the presented analyses, MRR data with $Z_e < -5$ dBz are discarded. This Z_e value corresponds to 0.02 to 0.09 mm/h depending on how Z_e varies with snowfall rate (see section 3.3).

To exclude rainfall events in Svalbard from the analysis, data from temperatures above -2°C were removed. Note that liquid precipitation events were used to verify the calibration of the MRR by comparing it with the reflectivity measured by an optical Parsivel disdrometer [Löffler-Mang and Joss, 2000]. No significant offset was found, so the calibration of the MRR in $\text{N}\ddot{\text{A}}$ is expected to be correct within ± 1 dB. No calibration reference is available for PE, but a calibration offset would only affect absolute values and not the profile structure.

3.3. Estimating Precipitation Rates From Radar Measurements

To convert the equivalent radar reflectivity factor Z_e into snowfall rates S , a power law relation is usually used:

$$Z_e = a \cdot S^b. \quad (1)$$

The parameters a and b depend on several assumptions, e.g., on the assumed particle habit, density, orientation, and particle size distribution. Backscattering properties of snow particles are increasingly affected by non-Rayleigh effects if the snowflake size is in the range of the radar wavelength. In addition, the natural variability of snowfall properties introduces a large uncertainty for any Z_e - S relation. Most Z_e - S relations for cloud radars have been derived for 35 GHz or 94 GHz, but not for 24 GHz. Calculations using an extended and updated version of the scattering database for snow particles [Liu, 2008b] revealed that the difference between backscattering at 24 GHz and at 35 GHz is below 0.5 dB, and hence, the 35 GHz Z_e - S relation is applied to MRR observations at 24 GHz in this study. Note that *Kneifel et al.* [2011] came to a different result because of interpolation effects across frequency: at that time, the database of *Liu* [2008b] did not include scattering estimations for 24 GHz.

To take the uncertainty of a Z_e - S relation into account, we use three Z_e - S relations from *Kulie and Bennartz* [2009], which are available for both 94 and 35 GHz (Table 2). These relations have been derived from aircraft measurements of particle size distribution and a large set of snow particle habits and their associated scattering properties. While the conversion of Z_e to S is necessary to illustrate the impact on snow amount at the surface, most analyses are performed in Z_e space in order to confine uncertainties to the vertical structure.

Note that due to increasing non-Rayleigh effects with increasing particle size, Z_e measured by MRR can be larger than that measured by CloudSat. To estimate this effect, a correction term

$$Z_{e(94\text{ GHz})} = \alpha \cdot Z_{e(35\text{ GHz})}^{\beta} \quad (2)$$

is derived by solving each pair of 35/94 GHz Z_e - S relations presented above for S (see Table 2). In Figure 3, the resulting differences for a Z_e range of -10 to 30 dBz are presented. It shows a significant decrease of Z_e with a high spread due to particle type. For example, an MRR measurement of ~ 10 dBz corresponds to CloudSat measurements between 1.4 and 7.7 dBz. In the following, the conversion for aggregates is used as an average of the different relations.

4. Comparison of MRR and CloudSat Above the Blind Zone

For our assumption that the MRR observations below the blind zone can be used to investigate the changes in Z_e within CloudSat's blind zone, we first have to investigate whether MRR generally reproduces the Z_e statistics measured by CloudSat above the blind zone. Because Z_e from both instruments cannot be directly

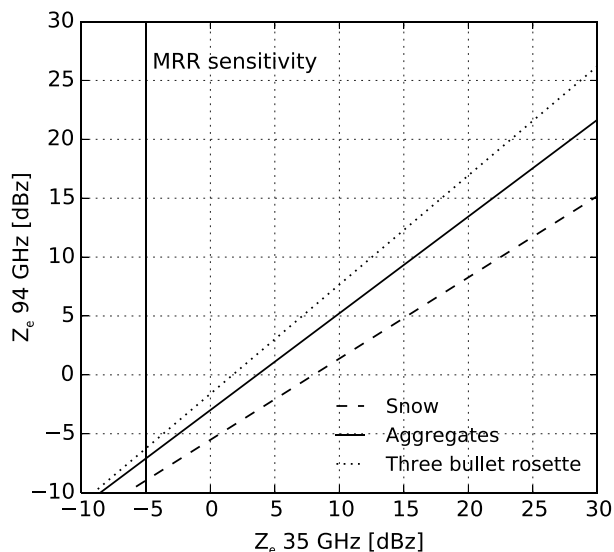


Figure 3. Conversion of reflectivity Z_e from 35 GHz to 94 GHz for snow (dashed), aggregates (solid), and three bullet rosettes (dotted) derived from the Z_e - S relations of *Kulie and Bennartz [2009]*.

compared due to the different radar frequencies, the comparison is performed in terms of S using the Z_e - S relations introduced in section 3.3.

For this, we also applied the MRR sensitivity threshold to CloudSat data. The -5 dBz threshold of the MRR corresponds to -7.10 dBz at 94 GHz assuming aggregates for equation (2) (-8.97 dBz for snow and -6.25 dBz for three bullet rosettes). In comparison to the snowfall threshold of -10 dBz defined by *Liu [2008a]*, this threshold reduced the number of events by 6 to 24%, depending on assumed particle type. Because these events only weakly contribute to total precipitation, snow mass is only reduced by 1 to 5%. Consequently, we can assume that the MRR is able to capture the majority of snowfall.

The comparison of the frequency distributions of derived snowfall rate S at H_{CS} for both PE and NÅ stations (Figure 4) reveals the generally strong similarity between CloudSat and MRR data at H_{CS} despite the different radar and data sampling characteristics. However, some differences between the CloudSat and MRR snowfall rate distribution occur particularly at PE, where CloudSat shows a broader frequency distribution shifted to higher S . Note that this discrepancy disappears if the MRR data are modified by an assumed calibration offset of $+2$ dB. Although we are currently unable to properly correct the calibration for the MRR at PE, such a constant bias would not affect the results of this study since we are using the same instrument for different heights and only analyze their relative differences. Attenuation effects might also lead to an offset between

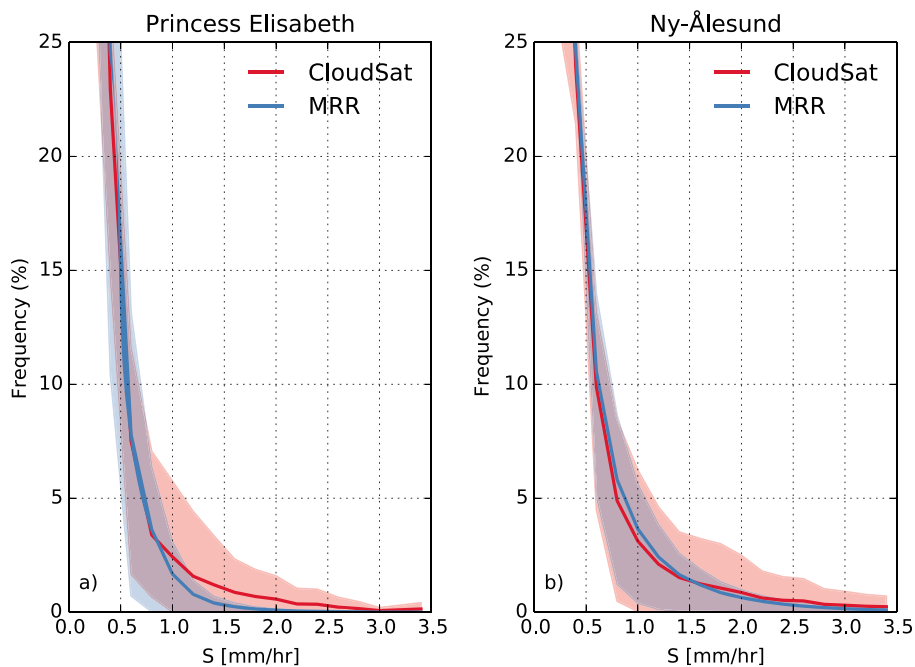


Figure 4. Frequency distribution of snowfall rates S for CloudSat (red) and MRR (blue) data at 1200 m agl (H_{CS}) for the (a) Princess Elisabeth and (b) Ny-Ålesund stations, where snowfall rate was determined by the range of Z_e - S relationships in *Kulie and Bennartz [2009]*. The line/polygon represents the mean/range of S for these Z_e - S relationships.

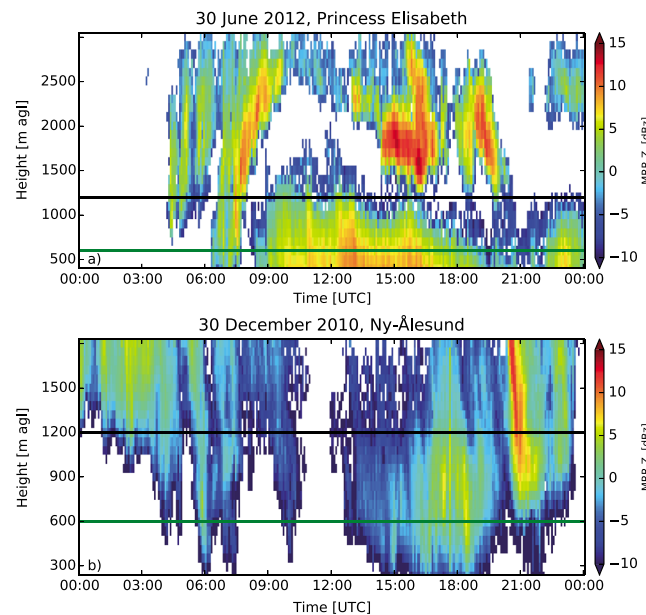


Figure 5. Time-height plots of reflectivity Z_e for two cases at (a) Princess Elisabeth and (b) Ny-Ålesund. CloudSat’s blind zone of 1200 m agl (H_{CS}) and a reduced blind zone of 600 m agl (H_{FM}) are denoted by a black and green line, respectively.

CloudSat and MRR because attenuation at 94 GHz is 1 order of magnitude larger than at 35 GHz, as shown by *Matrosov* [2007]. However, they found that attenuation is only relevant for higher-precipitation rates and thick snow layers, where attenuation can exceed 1 dB. In addition, attenuation is partly compensated for by multiple scattering effects in heavier snowfall [*Matrosov and Battaglia*, 2009]. Attenuation due to supercooled liquid water can be neglected at 24 GHz and is below 1 dB at 94 GHz for liquid water paths of less than 100 g m^{-2} [e.g., *Kneifel et al.*, 2014].

Further, the wide range of occurrences for a specific S (e.g., between 0 and 5% for $S = 1 \text{ mm h}^{-1}$ for CloudSat at PE) demonstrates that the uncertainty in S due to the use of different Z_e - S relationships associated with varying snow microphysical characteristics is much larger than the difference between MRR and CloudSat data.

5. Analysis of the Blind Zone

Based on our comparison of MRR and CloudSat at H_{CS} in the previous section, we can assume that the MRR observations within CloudSat’s blind zone can be used to investigate the impact of the blind zone on snowfall frequency and snowfall rate estimation. In the following, we will also include a blind zone height of 600 m ($H_{FM} = H_{\text{FutureMission}}$ hereafter) in the analysis. This height represents the most optimistic estimation of the blind zone of future satellite missions.

The different possible errors introduced by assigning the reflectivities at H_{CS} to the surface level can be identified in Figure 5 showing example days of the observed, rather complex but typical, precipitation structures. In the example at PE (Figure 5a), two precipitation layers can be identified during the day. They appear to be disconnected when observed with an MRR: a high-precipitation layer with high Z_e values does not necessarily imply that precipitation reaches the ground. For the NÅ case (Figure 5b), the effect of sublimating precipitation or virga is even more pronounced with precipitation sometimes reaching surface levels but most of the time sublimating aloft. Consequently, in these cases, the blind zone would lead to an overestimation of snowfall on the surface (hereafter referred to as *commission error*). At PE, the opposite effect can also be seen: shallow precipitation that would be potentially missed by a satellite radar and thus resulting in an underestimation of snowfall on the ground (hereafter called *omission error*).

In the following subsections, we analyze the ground-based data sets focusing on different aspects: what is the impact of the blind zone on the climatology of observations in respect to (1) Z_e distribution, (2) number of events, and (3) single events? Finally, we discuss the effect of the blind zone by considering precipitation amount.

5.1. Impact on Z_e Distribution

In the following, we investigate whether the vertical distribution of Z_e is affected by virga, shallow precipitation or microphysical processes such as sublimation, riming, and aggregation. For this investigation, reflectivity versus altitude 2-D histograms (2-DH) of observed reflectivities for the complete observation period of 1 year are shown in Figures 6b and 7b for PE and NÅ, respectively. To enable a better interpretation of the MRR statistics in terms of CloudSat observations, results are shown for both 24 GHz Z_e and converted to 94 GHz assuming aggregates (see Figure 3 for the impact of assuming other snow types). For both sites, the 2-DH is very homogeneous between H_{SF} and H_{CS} . The median of reflectivity is constant with altitude

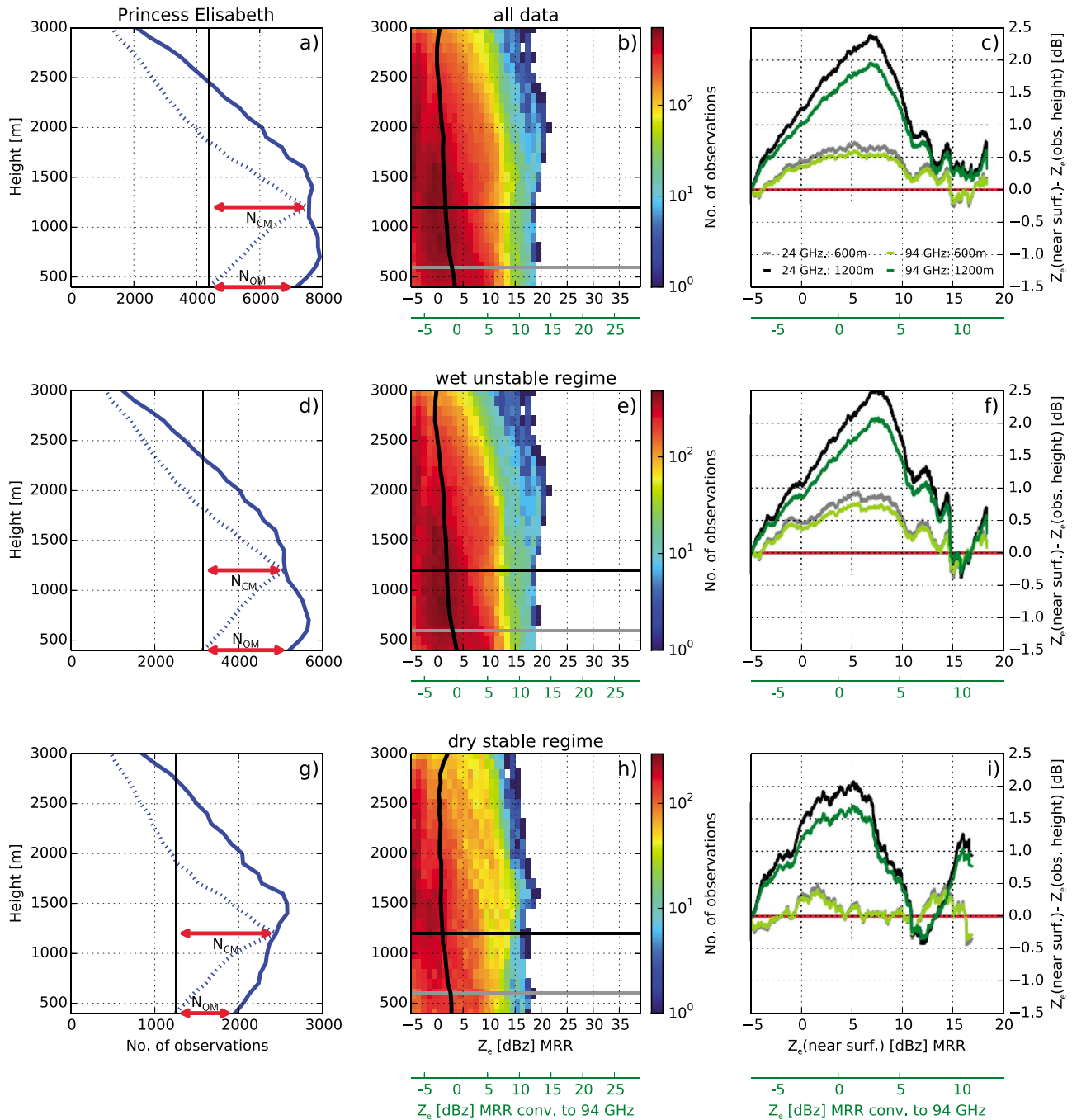


Figure 6. (a) Total number of observations N with Z_e of MRR observations at Princess Elisabeth station, East Antarctica, larger than -5 dBz (solid blue line) are compared with profiles \hat{N} , which also contain snowfall at 1200 m agl (dashed blue line) for the complete data set. Commission and omission errors (N_{CM} and N_{OM}) are marked with red arrows. (b) Reflectivity versus altitude 2-D histograms (2-DH) of observed MRR reflectivities. The median profile is denoted by the black solid line. The horizontal, black line denotes H_{CS} . A reduced blind zone of 600 m agl (H_{FM}) of a future satellite mission is marked with a horizontal, gray line. An estimate for the corresponding Z_e at 94 GHz using the coefficients for aggregates is indicated by the additional, green scale. (c) Detrended Quantile-Quantile (DQQ) plots of the reflectivity observations close to the surface (H_{SF}) in comparison to H_{CS} (black) and in comparison to H_{FM} (gray) as well as after conversion to 94 GHz (green and light green lines). In contrast to a Quantile-Quantile plot, only the differences between quantiles of Z_e at H_{CS} (or H_{FM}) and H_{SF} are shown for the ordinate, i.e., a value of zero means perfect agreement. (d–f) Same as Figures 6a–6c, but only for the *wet unstable regime*. (g–i) Same as Figures 6a–6c, but only for the *dry stable regime*.

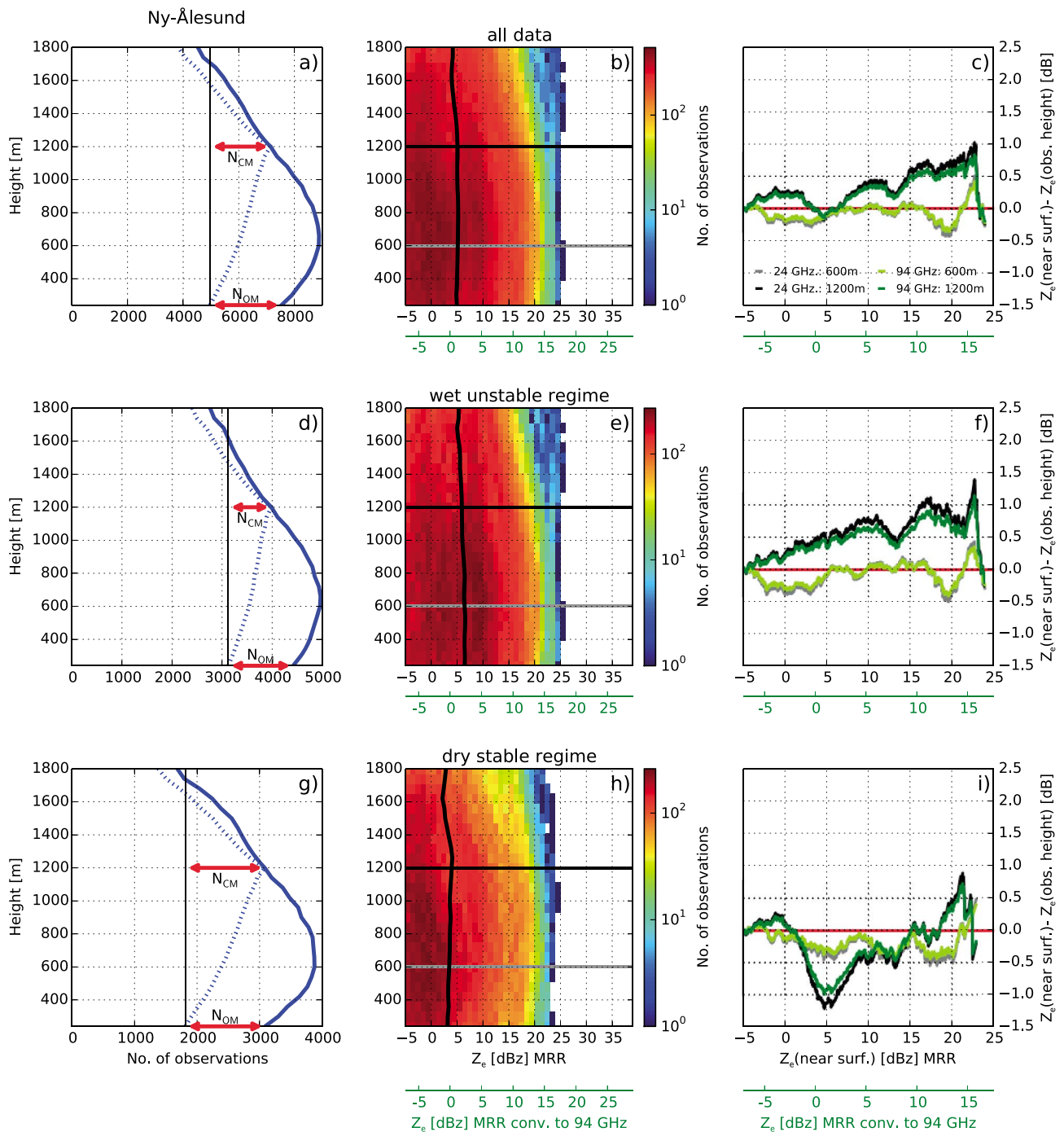


Figure 7. Same as Figure 6 but for Ny-Ålesund, Svalbard.

in $N\ddot{A}$ and shows only a weak increase of 1–2 dB toward the surface at PE. This low-reflectivity gradient is highly consistent with the findings of an almost constant Z_e profile found by *Henson et al.* [2011] for the Canadian Arctic.

The distribution of reflectivity measurements at H_{CS} and H_{SF} is also compared using Detrended Quantile-Quantile (DQQ) [Thode, 2002] plots (Figures 6c and 7c). For this, both series of Z_e are divided into n_z quantiles where n_z is the length of the shorter Z_e series. Then, the quantiles are sorted and the quantiles of H_{CS} and H_{SF} are subtracted from each other. This difference of quantiles is plotted against the quantiles for Z_e at H_{SF} . By this, the distribution of Z_e at H_{CS} can be directly compared to the distribution at H_{SF} and even small

differences between these distributions become visible. In contrast to the analysis of vertical gradients of individual profiles, this method has the advantage that it is not influenced by changes of the Z_e profile due to, e.g., advection or temporal evolution. For PE, the distribution of Z_e at H_{CS} is shifted toward smaller reflectivities by a maximum of 2.3 dB in the medium Z_e range. The agreement of distributions increases for Z_e values larger than 10 dBz, indicating more similar distributions for stronger events. However, due to the relatively small number of events in this high-reflectivity region, a more extended data set would be necessary for a clear conclusion. For $N\dot{A}$, agreement of Z_e distributions is better, and the shift between distributions at H_{CS} and H_{SF} is mostly below 1 dB, often below 0.5 dB with a slight increase toward larger Z_e . Performing the comparison with MRR data converted to 94 GHz has only minor impact on the results.

To investigate whether the found changes in statistics depend on the regime classification, the analysis is repeated for the two synoptic regimes. For the *wet unstable regime* (Figures 6e, 6f, 7e, and 7f), only small differences to the complete data set are visible in the 2-DH and the DQQ plot. However, the *dry stable regime* (Figures 6h, 6i, 7h, and 6i) features, on average, lower reflectivities for both sites, especially at the levels below H_{CS} because moisture supply for particle growth in the near-surface layers is low while temperatures are relatively high, favoring particle sublimation. In $N\dot{A}$, the distribution of Z_e at H_{CS} is shifted by up to -1 dB in comparison to HSF for the *dry stable regime*, which is probably related to sublimation effects (Figure 7i). For PE, the shift of the distribution is similar to the one using the complete data set—except for rare values exceeding 10 dBz. Sublimation effects lead to a reduction of the shift of Z_e distributions by 0.3 dB (Figure 6i).

To investigate the impact of a reduced blind zone, DQQ plots are also presented for an observation height of 600 m agl (H_{FM} , Figures 6c, 6f, 6i, 7c, 7f, and 7i). For $N\dot{A}$, this results in a significantly better statistical agreement of H_{FM} and H_{SF} . The offset is only ± 0.5 dB for all regimes. For PE, the agreement of Z_e distributions is improved and the offset of the distributions is below ± 1 dB and only for the *dry stable regime* (Figure 6i) is the offset negligible and below ± 0.5 dB. In summary, a reduction of the blind zone by 50% significantly improves the agreement of the Z_e distributions at H_{SF} and the observation altitude. We expected the impact on Z_e distribution to still be underestimated for H_{CS} and H_{FM} at PE because H_{SF} is at an altitude of 400 m agl and processes between the surface and H_{SF} are not considered. In $N\dot{A}$, H_{SF} is at 240 m agl, so the underestimation is expected to be less. The results for LY, which are very similar to $N\dot{A}$, can be found in Figure S1 in the supporting information.

5.2. Impact on the Number of Snowfall Events

To investigate the impact of the blind zone on the total number of observed precipitation events, N —defined as the number of observations greater than -5 dBz—is calculated as a function of height (Figures 6a and 7a). By restricting the number of events in one height to those with snowfall at H_{CS} (\hat{N} in Figures 6a and 7a), the number of shallow events (omission error, N_{OM}) can be estimated by the difference between N and \hat{N} . At PE ($N\dot{A}$), 38.7% (34.1%) of 7153 (7537) observations are classified as omission errors because they are present at H_{SF} but not at H_{CS} above. For PE, this number might even be too low if shallow events occur below H_{SF} of 400 m agl.

Virga events that sublimate and do not reach H_{SF} , although they are present at H_{CS} (commission error, N_{CM}), can be estimated by the decrease of \hat{N} relative to its maximum at H_{CS} . As a result, 44.4% (28.9%) of the events observed at H_{CS} do not reach the surface at PE (in $N\dot{A}$). So N_{OM} is similar for PE and $N\dot{A}$, while N_{CM} is not.

It is important to note that N_{OM} and N_{CM} not only contain shallow and sublimating precipitation events but are also affected by advection and shear effects as well as by the terminal fall velocity of snow, which also causes slanting of the observed snowfall profiles. When a cloud starts to precipitate, particles need up to 20 min from H_{CS} to the surface assuming a mean fall velocity for snow of 1 m s^{-1} . During these first 20 min, the event will be classified as a commission error. On the other hand, if the cloud stops precipitating, the last 20 min of the event will be classified as an omission error. Although N_{CM} and N_{OM} might be overestimated, they are affected in the same way in the long term. Thus, the difference between both errors is expected not to be biased by the slanting of the profiles. If advection of precipitation is assumed to be homogeneously distributed, advection effects are also expected to cancel each other out.

The difference between N_{OM} and N_{CM} is equal to the difference of N between the altitudes at H_{CS} and H_{SF} . For PE, N is reduced by 5.7% from H_{CS} toward the surface (Figure 6a). In contrast, N increases by 5.2% at the surface for $N\dot{A}$ (Figure 7a). Interestingly, the increase toward the surface is not monotonic, but the vertical distribution of N is “belly shaped” with a maximum at H_{FM} . Since the decrease toward H_{SF} is smaller than the

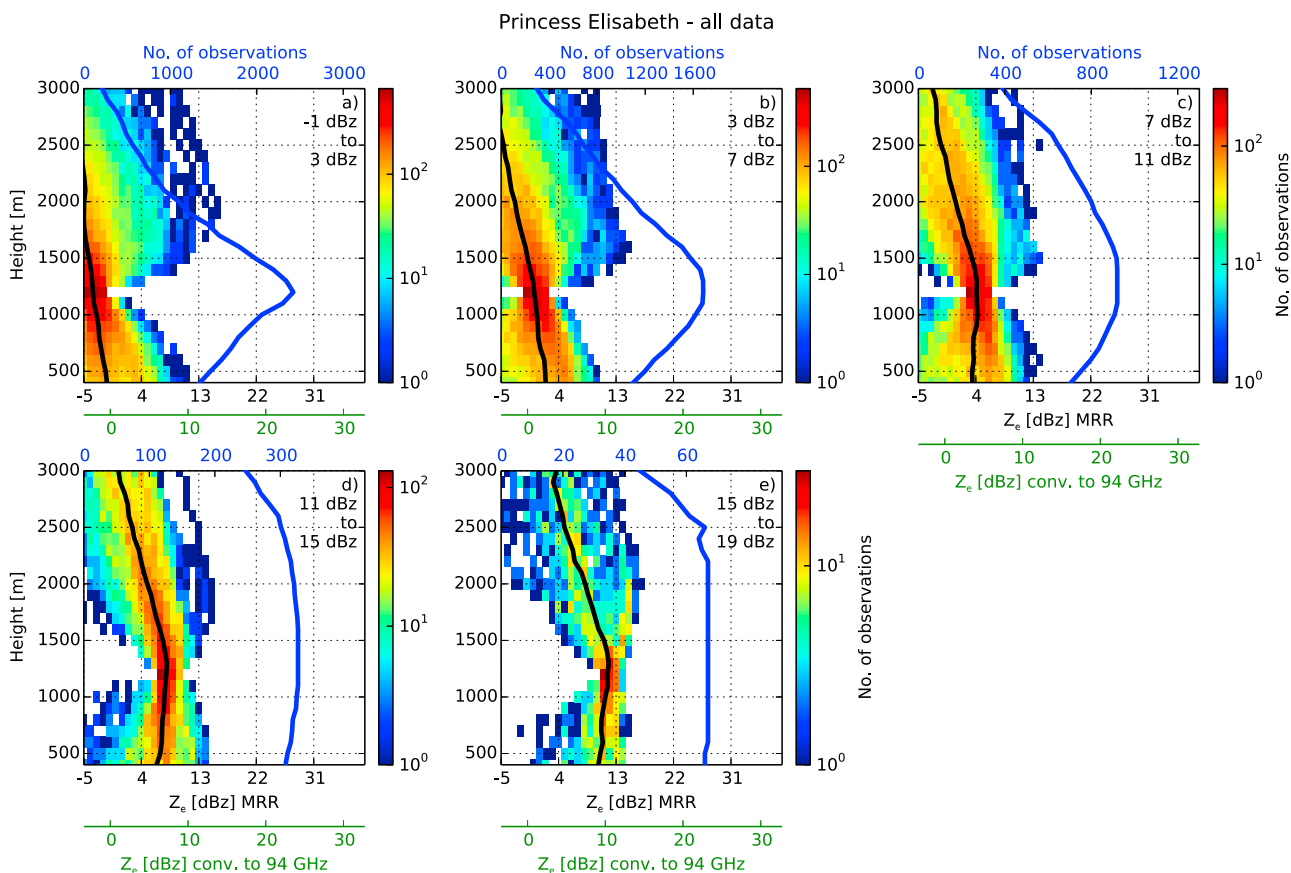


Figure 8. Reflectivity versus altitude 2-D histogram (2-DH) of MRR observations at Princess Elisabeth station, East Antarctica. The presented data set is limited to profiles which are within a certain reflectivity threshold at the CloudSat reference height of 1200 m agl (H_{CS}). (a) Only profiles that are between -1 and 3 dBz at the reference height. (b–e) Limited to reflectivities of -5 , 9 , 13 , and 17 ± 2 dBz, respectively. The median is denoted in black, the total number of observation greater -5 dBz per height in blue. An estimate for the corresponding Z_e at 94 GHz using the coefficients for aggregates is indicated by the additional, green scale.

increase to H_{CS} , a net increase of N occurs at $N\hat{A}$. In other words, at PE, virga events are more frequent than shallow precipitation (Figure 6a), whereas at $N\hat{A}$, shallow precipitation occurs more frequently (Figure 7a).

In $N\hat{A}$, the maximum occurs at H_{FM} , which implies that a reduced blind zone with H_{FM} would lead from an underestimation of N to an overestimation by 17.8%. For PE, the overestimation of N would only slightly change from 5.7% to 8.5%. This deterioration is in contrast to the finding that the statistical representation of Z_e is improved for a lower blind zone (Figures 6c and 7c). Apparently, the combination of virga and shallow precipitation leads to a local maximum of occurrence between H_{FM} and H_{CS} .

While there is a clear tendency toward larger N below H_{CS} for the *wet unstable regime* indicating more frequent shallow precipitation (Figures 6d and 7d), a stronger decrease of snow events can be observed for the *dry stable regime* at both sites (but more pronounced at PE, Figures 6g and 7g). This stronger decrease might be attributable to strong sublimation and virga formation.

5.3. Impact on Individual Snowfall Events

While the previous sections investigated the mean occurrence of snow characteristics, we now highlight the large spread if single events are considered. This procedure is important because the 16 day repeat cycle of CloudSat means that a particular precipitation event is usually only observed once and temporal evolution cannot be measured. For this investigation, we sort the profiles with respect to their Z_e values at H_{CS} ; the resulting 2-DH diagrams for the different Z_e intervals at H_{CS} are shown in Figure 8 for PE and Figure 9 for $N\hat{A}$ for the original measurement and also converted to 94 GHz. While events with Z_e of up to 7 dBz at H_{CS} tend to increase toward the surface on average by 0.5 to 3 dB, events with larger Z_e decrease by up to 3 dB. Transferred to a CloudSat observation, this result means that CloudSat would tend to underestimate weaker

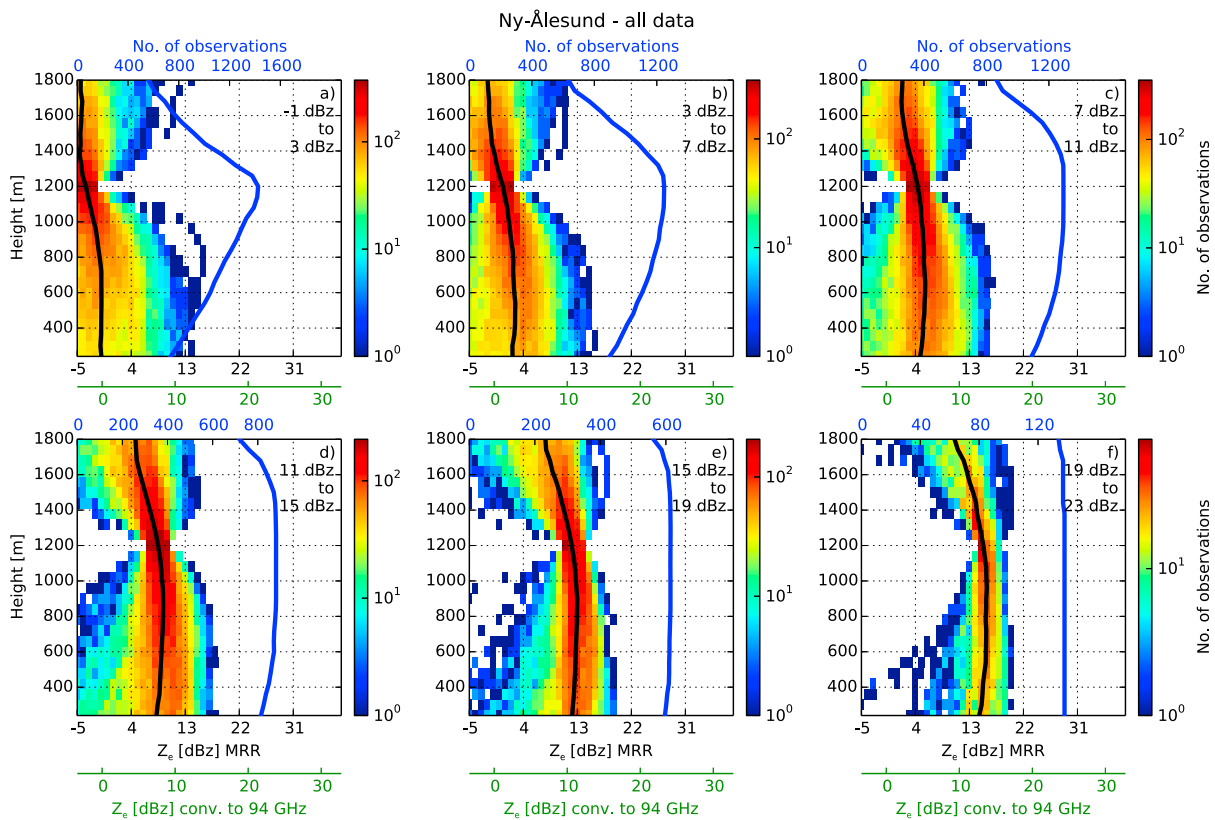


Figure 9. (a–e) Same as Figure 8 but for Ny-Ålesund with (f) an additional panel for the 19 to 23 dBz interval.

precipitation events while it would overestimate stronger events. This finding also holds for LY, and the corresponding plot can be found in the supporting information (Figure S2).

While for lower Z_e up to 3 dBz, this finding is consistent with our findings in section 5.1, for higher Z_e , an analysis of the DQQ plots revealed a shift of the Z_e distribution to higher reflectivities at H_{CS} in comparison to at H_{SF} (Figures 6c and 7c). This result highlights the vertical inhomogeneity of precipitation events at a single time step, and it can be also seen in the large spread of observations at H_{SF} for a single Z_e interval being larger for NÅ than for PE.

5.4. Impact on the Total Mass Flux

The hydrological cycle is driven by the flux of precipitation to the surface. To investigate the impact of the blind zone on mass flux, Figure 10 presents an estimation of the snowfall rate profile for PE and NÅ. The figure reveals how the different reflectivity values contribute to the total precipitation as a function of height using the Z_e - S relation for snow presented in section 3.3. The total liquid equivalent snowfall amount of the 12 month data set derived from MRR is roughly 200 mm at PE and 320 mm in NÅ, although uncertainty of the absolute values is large due to the application of the Z_e - S relation. Hence, we normalized the precipitation amount at every height with respect to the total precipitation amount at H_{SF} , resulting in a reduced sensitivity of the precipitation amount profile to uncertainties in the Z_e - S relation.

The vertical changes of Z_e are a combination of changes in the reflectivity distribution (section 5.1) and the variation of N with height (section 5.2). In general, events with reflectivities between 0 and 10 dBz (corresponding to 0.2 and 0.7 mm/h, respectively, using the Z_e - S relation for snow) contribute most to the total precipitation amount at PE (Figure 10a), underlining the need for a high radar sensitivity. This finding is in agreement with the results from *Kulie and Bennartz [2009]*, who found that most precipitation in the northern periphery of Antarctica originates from precipitation rates of 0.2 to 1 mm/h and indicates that our data set is typical for Antarctica. Events with reflectivities larger than 15 dBz are so rare that they hardly contribute to the total precipitation amount. The contribution of the various classes of reflectivity is different for

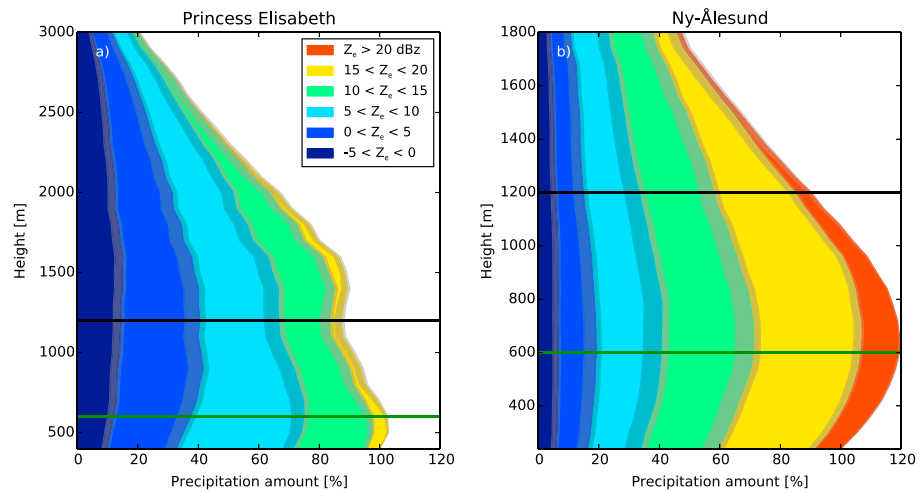


Figure 10. Contribution of various reflectivity intervals to the total precipitation amount in dependence on height for (a) Princess Elisabeth and (b) Ny-Ålesund. For the colored areas, the Z_e - S relation by *Kulie and Bennartz* [2009] for snow is used. Uncertainty of the borders between the different intervals due to the Z_e - S relations is estimated by the gray, shaded area, which is estimated by applying also Z_e - S relations for *three bullet rosettes and aggregates* by *Kulie and Bennartz* [2009]. The figures are normalized by total surface precipitation. CloudSat’s blind zone of 1200 m agl (H_{CS}) and a reduced blind zone of 600 m agl (H_{FM}) are denoted by black and green lines, respectively.

NÅ (Figure 10b): most precipitation originates from events with reflectivities between 15 and 20 dBz, which is related to the, on average, warmer and moister climate in Svalbard.

The vertical development of the different reflectivity classes shows how behavior is different for NÅ and PE. For PE, the contribution of the 0 to 5 dBz interval and the 10 to 15 dBz interval changes little for the height between H_{CS} and H_{SF} . The contribution of the 5 to 10 dBz interval, however, increases toward the surface. At PE, the total mass flux at H_{CS} is underestimated by 11 percentage points of total precipitation in comparison to H_{SF} . At H_{FM} , this changes to an overestimation of 3 percentage points, but uncertainty is high because, for PE, observations below 400 m are not available and the further development toward the ground is unknown.

In NÅ, the contribution of the three intervals between 5 and 20 dBz is even more similar at H_{CS} and H_{SF} . For the range between H_{CS} and H_{SF} , however, all reflectivity intervals larger than 5 dBz contribute more to the total precipitation than at H_{CS} and H_{SF} . Apparently, omission and commission errors are competing but cancel out each other close to the surface. This result would also mean that a future satellite-based radar mission with a smaller blind zone would not improve the estimation of the total mass flux. Instead, at least for NÅ, total mass flux would be overestimated by 19 percentage points—in comparison to 9 percentage points underestimation at H_{CS} . The results are similar for LY (see Figure S3 in the supporting information) and prove that the belly shaped distribution is not a local effect.

These results are robust as demonstrated by using again the two other Z_e - S relations introduced in section 3.3 to estimate the uncertainty. As can be seen from Figure 10, applying a different Z_e - S relation has only a minor effect on the results and does not change the overall shape of the distribution. Even if the Z_e - S relation changes with height, the impact on the total precipitation distribution is less than 5%.

6. Summary and Conclusion

The impact of CloudSat’s blind zone below 1200 m agl (H_{CS}) on snowfall statistics was investigated for three polar sites, the Belgian Princess Elisabeth station (PE) in East Antarctica and for Ny-Ålesund (NÅ) and Longyearbyen (LY, see supporting information for results) in Svalbard, Norway (Figure 1). To investigate the impact on a future satellite mission with a reduced blind zone, a blind zone of 600 m agl (H_{FM}) was also considered.

We used a ground-based Micro Rain Radar and compared with the lowest clutter-free CloudSat observation height H_{CS} (Figure 4). The MRR was found suitable because differences due to the different footprint

and temporal resolution were found to be less than the uncertainty caused by the applied Z_e - S relation. Consequently, we investigated the blind zone effects by comparing MRR observations at H_{CS} with MRR observations near surface (H_{SF}), assuming that MRR observations are representative of CloudSat observations. To our knowledge, the 12 month MRR data sets from NÄ and LY are the first precipitation radar observations available for Svalbard. The MRR at PE is currently the only ground-based precipitation radar in Antarctica.

The frequency distribution of MRR reflectivity changes between H_{CS} and H_{SF} . For PE (Figure 6c), the distribution of Z_e is shifted by up to 2.5 dB toward smaller values if measured at H_{CS} . For NÄ (Figure 7c), the shift is much smaller and below 1 dB. A reduction of the blind zone by 50% leads to a reduction of the offset by more than a factor of 2 for PE, and in NÄ the remaining shift is negligible.

The better agreement due to a reduction of the blind zone cannot be seen when analyzing the total number of observations N (Figures 6a and 7a): at H_{CS} , N is overestimated by 5.7% and underestimated by 5.2% for PE and NÄ, respectively. At H_{FM} , this result changes to overestimations of N of 8.5 and 17.8%. This change is more strongly pronounced at NÄ and most likely due to competing processes: virga and shallow precipitation.

Based on temperature inversion and relative humidity, the data set was divided into two regimes: the *dry stable* and the *wet unstable regime*, with the latter representing 73% and 59% of all precipitation events at PE and NÄ, respectively. For the *dry stable regime*, the overestimation of N at H_{CS} and H_{FM} is most pronounced and Z_e decreases toward the ground, i.e., virga effects dominate, as might be expected due to increased sublimation. For the *wet unstable regime*, agreement of N at the different levels is better. This indicates that for these regimes, which are more related to the influence of synoptic disturbances, the profiles are on average slightly more constant in the vertical. This classification might be exploited for mitigation strategies.

Agreement of observations between H_{CS} and H_{SF} is less when single events are considered (Figures 8 and 9) because precipitation intensity strongly varies with height due to the fall velocity of ice particles and due to advection effects. Hence, low-precipitation intensity gets generally enhanced and high-precipitation intensity becomes generally weaker.

The change in both N and Z_e contributes to the estimation of precipitation amount (Figure 10). For NÄ, the belly shape of N can be also seen in the total precipitation amount: at H_{CS} , total precipitation is underestimated by 9 percentage points, while at H_{FM} , it would be overestimated by 19 percentage points, which is mainly driven by events with $Z_e > 15$ dBz. For PE, such a belly shape cannot be clearly seen, but the underestimation of total precipitation of 11 percentage points at H_{CS} still changes to an overestimation of 3 percentage points at H_{FM} showing that virga and shallow precipitation effects are also partly overlapping at PE.

In general, our results are only valid for the vicinity around the station. However, due to the spatial homogeneity of surface properties in East Antarctica, we expect our findings to be representative of a much larger area. For Svalbard, orography is much more complex, limiting the spatial representativeness of this study. Because results for an identical setup in LY, located 110 km southeast of NÄ, are comparable (see supporting information), we are nevertheless confident that our results are representative at least for the western region of Svalbard.

In summary, shallow precipitation and virga effects are found to compete and lead to a change in the number of observed events and total mass flux of 9 to 11 percentage points, although the statistical distribution of Z_e changes only little. While the statistical agreement is enhanced with a potentially lower blind zone altitude, the difference in the number of observed events and total mass flux increases. Hence, a reduced blind zone would not improve snowfall observations in all aspects.

For the future, more radar observations would allow further investigation of this issue, also in other regions. Since heavy snowfall events are rare but contribute strongly to the total precipitation, longer time series are needed. The question of how total precipitation translates into snow accumulation at the surface is complex [Gorodetskaya et al., 2014a, 2014b] and needs to be studied in more detail in the future. The MRR comprises a low-cost, low-maintenance alternative to more complex systems and has been successfully operated at PE during the unmanned winter period. It thus offers potential for investigating snowfall in other polar regions. Special care should be taken to configure the radar with a lower minimum

observation height, allowing observations as close to the surface as possible. Also, the use of a radar with a higher sensitivity is desirable so that the impact of the blind zone on events with reflectivities below -5 dBZ can be investigated. For this, data from the North Slope of Alaska site of the Atmospheric Radiation Measurement Program in Barrow [Mather and Voyles, 2012], from the Summit Station in Greenland [Shupe et al., 2013] or from the Zugspitze in the German Alps [Löhnert et al., 2011], as well as possibly from other sites, could be used for future studies.

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