



Precipitation processes in an Antarctic moist air intrusion: insights from multi-frequency radar observations over a 1100-km transect

Heather Corden¹, Julien Delanoë², Massimo Del Guasta³, and Alexis Berne¹

¹Environmental remote sensing laboratory, École Polytechnique Fédérale de Lausanne (EPFL), Lausanne, Switzerland

²LATMOS/IPSL, UVSQ Université Paris-Saclay, Sorbonne Université, CNRS, Guyancourt, France

³Istituto Nazionale Ottica CNR, Sesto Fiorentino 50019, Italy

Correspondence: Heather Corden (heather.corden@epfl.ch) and Alexis Berne (alexis.berne@epfl.ch)

Abstract. In Antarctica, intrusions of coastal moist and warm air onto the high plateau play an important role in the mass balance of the ice sheet, due to their significant contribution to annual snowfall accumulation. The synoptic drivers of warm intrusions are well-established, in particular for extreme cases linked to atmospheric rivers. However, a lack of suitable observations means the micro-physical cloud and precipitation processes within intrusions remain uncertain. In the context of the Atmospheric Water Cycle over Antarctica (AWACA) project, we investigate an intrusion associated with a coastal cyclone in East Antarctica in February 2025. Lagrangian trajectory analysis confirms that air masses within the intrusion pass over a 1100-km observational transect from the coast to the plateau, allowing precipitation properties to be tracked as the intrusion moves inland. At four sites along the transect, a multi-frequency, polarimetric, spectral radar dataset is used to investigate micro-physical processes. The reflectivity and dual frequency ratios indicate a decrease in particle size as the intrusion moves inland. Near the coast, high fall-velocities and spectral signatures point to riming of snowflakes, fuelled by ascending air masses above the coastal slope and the availability of supercooled water. On the plateau, dry and cold conditions lead to smaller particles, for which variation in the radar signal appears to arise from primary ice crystal habits. The case study illustrates the potential of multi-frequency radar data from an autonomous observational transect to investigate precipitation processes between the Antarctic coast and plateau.

1 Introduction

Intrusions of warm moist air masses from the coast onto the Antarctic plateau have a large impact on the surface mass balance of the Antarctic ice sheet. These events can lead to enhanced snowfall (Wille et al., 2021), high winds (Turner et al., 2009), rare liquid precipitation and surface melting (Gorodetskaya et al., 2013; Wille et al., 2025). Here, we focus on the clouds and precipitation associated with such warm intrusions. Extreme snowfall events, many of which are associated with warm intrusions, are known to contribute significantly to annual accumulation totals in Antarctica: snowfall from the 10% of days with the largest accumulation can contribute up to 40% of annual accumulation, with an increasing proportion close to the coast (Turner et al., 2019). The strongest warm intrusions may be associated with so-called atmospheric rivers (AR) – extended filaments of enhanced moisture transport – which have been a focus of recent research (Wille et al., 2025, and references therein). For example, MacLennan et al. (2022) calculated that ARs contribute 13% to annual Antarctic snowfall, a value



25 known to vary with the AR detection method used but believed to be conservative. Indeed, using a new 3D method, Takahashi
et al. (2026) found a higher contribution of 50-80% for East Antarctica. On a more local scale, Gorodetskaya et al. (2014)
found that high-accumulation years at Princess Elisabeth station (PE) were those for which a greater number of ARs reached
the Antarctic coast. The frequency and intensity of ARs reaching Antarctica are expected to increase in a warming climate
(O'Brien et al., 2022; Wille et al., 2025). The significance of coastal-driven moist intrusions towards present and future annual
30 precipitation totals motivates the investigation of the precipitation processes within these systems. In particular, the response of
snowfall-microphysics within intrusions to the changing atmospheric conditions – temperature, humidity and wind – between
the coast and the high Antarctic plateau remains unclear.

Ground-based meteorological radars offer unique insights into the macro- and micro-physical properties of liquid and solid
precipitation. Time-height profiles of the radar equivalent reflectivity factor (Z_e or reflectivity for short) and mean Doppler
35 velocity (MDV) provide information on the size, number and fall velocity of particles. This can be used to estimate snowfall
fluxes (eg. Matrosov et al., 2009; Souverijns et al., 2017; Grazioli et al., 2017a; Scarchilli et al., 2020) and to reveal macro-
physical effects such as snowfall sublimation in dry katabatic layers (Grazioli et al., 2017b; Scarchilli et al., 2020; Gehring
et al., 2022). It can also be used to infer micro-physical processes, such as the rapid increase in fall speed due to riming
(eg. Grazioli et al., 2015; Kneifel and Moisseev, 2020). When the full reflectivity-weighted spectrum of Doppler velocities
40 at each range gate (Doppler spectrum) is recorded, properties of the fastest and slowest falling particles can be differentiated.
Multi-modality of the spectrum can indicate the presence of different particle populations with distinct properties. For example,
multi-modality is expected when secondary ice production (SIP) – such as collisional breakup of larger snowflakes – leads to
the formation of a new population of small ice particles (Field et al., 2017; Billault-Roux et al., 2023; Kalesse et al., 2016).

The addition of polarimetric information from a second receiver channel provides information about the shape and ho-
45 mogeneity of the particles. In particular, the linear depolarisation ratio (LDR, the ratio of cross-polar reflectivity to co-polar
reflectivity when only one channel is transmitting) is lower for spherical particles (Ryzhkov and Zrnica, 2019). When combined
with the Doppler spectrum, the spectral LDR can aid in the identification of particle populations in the Doppler spectrum, such
as the low LDR of spherical super-cooled water droplets (Billault-Roux et al., 2023). Finally, the use of multiple radars at
different frequencies allows the Dual Frequency Ratio (DFR, also referred to as the dual wavelength ratio) to be calculated: the
50 ratio of reflectivities measured at different frequencies. The DFR can be used to identify regions of particle growth: as larger
particles exit the Rayleigh scattering regime of the smaller-wavelength radar, the rate of increase of reflectivity with particle
size decreases and the difference in reflectivity between the two wavelengths increases (eg. Matrosov et al., 2005; Moisseev
et al., 2015). Adding a third frequency can reveal signatures of particle density and help differentiate between snowfall growth
through riming and aggregation (Kneifel et al., 2015; Leinonen and Moisseev, 2015).

55 Despite the considerable logistical challenges, the value of meteorological radar data for understanding cloud and precipi-
tation processes has led to a growing number of measurement campaigns in Antarctica (Grazioli et al., 2023, and references
therein). For example, at Dumont d'Urville station (DDU), two radars were deployed for the austral summer of 2015-16 (Grazi-
oli et al., 2017a), of which the K-band Micro Rain Radar (MRR, see section 2.1) was maintained for an extended measurement
series (Wiener et al., 2024). Observations outside the austral summer season have historically been limited to the use of the



robust MRR system (Durán-Alarcón et al., 2019; Alexander et al., 2023), although more recent campaigns at McMurdo and Neumayer stations have introduced year-round operation of more complex radars alongside lidars and microwave radiometers (Lubin et al., 2020; Radenz et al., 2024).

The Atmospheric Water Cycle over Antarctica (AWACA) project includes multi-frequency, polarimetric and Doppler-spectra radar observations along a 1100-km transect in East Antarctica, with autonomous operation beyond the summer season. It therefore provides a novel opportunity to investigate the differences in precipitation processes between the coast and the high plateau. This paper presents a case study of a moist intrusion from the first months of operation. The case study was chosen due to the alignment of air trajectories over the observational transect (presented in section 3.1). The alignment means that it is possible, for the first time, to track the evolution of snowfall properties and processes from ground-based radar observations as the intrusion moves inland.

Section 2 introduces the AWACA transect and the radar data used in the case study. After a synoptic and macro-physical overview in Section 3, we track the micro-physical processes within the warm intrusion as it travels over each site of the observational transect (Sections 4-8). Section 9 provides a discussion and summary of the case study.

2 Data and instrumentation

2.1 Radars of the AWACA transect

The AWACA transect extends from DDU (66.66° S, 140.00° E, 40 m a.s.l.) on the coast to Concordia station (75.10° S, 123.35° E, 3233 m a.s.l.) on the Antarctic plateau. At DDU, instruments requiring higher power and monitoring by the wintering personnel are installed, though these are not used in this study. The remainder of the transect consists of four container-stations, known as Operational Platform Units (OPUs). The locations are shown in Fig. 1 and Table 1. Three of the OPUs (D17, D47 and D85) include their own energy system and are not accessible outside the summer season. The installation is designed for autonomous operation with remote monitoring through satellite internet. The final OPU is located at Concordia station (DMC) but contains the same instruments as the remote sites. Note that the names of the sites have no scientific significance and refer only to consecutively numbered points recognised by the French Polar Institute. The transect lies on the route used by the terrestrial convoy which transports supplies from DDU to DMC. Importantly, it is also approximately aligned with the climatology of air mass trajectories which lead to precipitation at Concordia, allowing the transformation of these air masses between the coast and plateau to be investigated. The OPUs have a smaller spacing near the coast in order to capture the higher gradient in atmospheric conditions in this region. The transect was deployed in the austral summer season from December 2024 to January 2025 and will remain in place for three years.

Figure 2 shows the instrumentation of the OPUs. Each OPU includes, among other instruments, three vertically-pointing radars with Doppler spectrum capabilities: a K-band Micro Rain Radar (MRR-Pro, or MRR hereinafter) and K_a-band MIRA cloud radar from Metek GmbH, and a W-band BASTA cloud radar developed by LATMOS (Delanoë et al., 2016). The radar characteristics are given in Table 2.

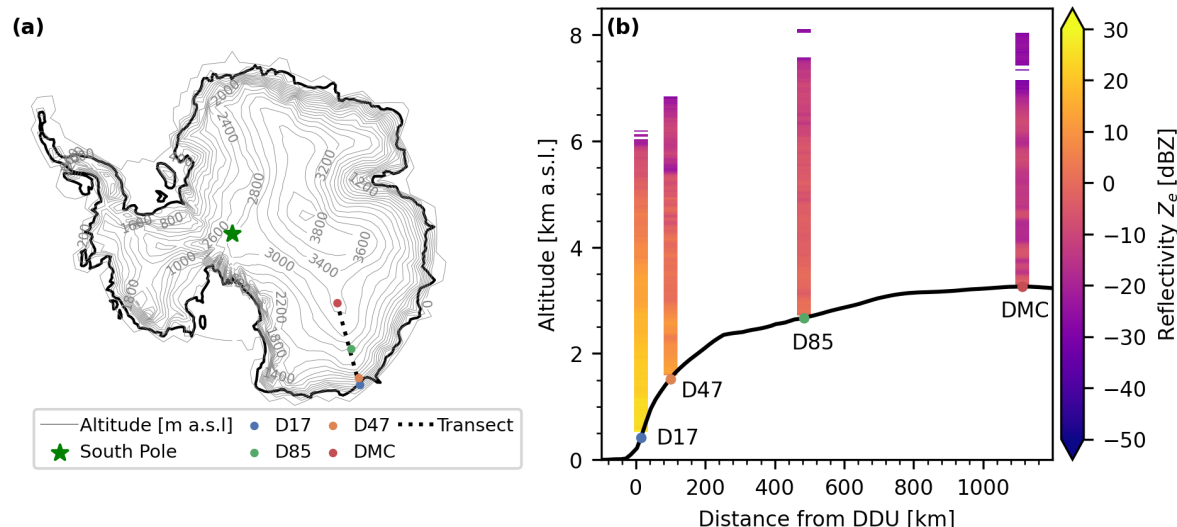


Figure 1. The AWACA transect. (a) shows the location of the four Operational Platform Units (OPUs), adapted from (Wiener et al., 2024). The coordinates of the sites are given in Table 1. The altitude contours (light grey) are taken from (Schaffer et al., 2016). (b) shows a vertical cross-section of the transect from Dumont d'Urville station (DDU) to Concordia station (DMC). The equivalent reflectivity measured by the MIRA radars at the four OPUs is shown for 06:00 UTC on the 17th of February 2025, after the arrival of the moist intrusion at all sites.

Table 1. Coordinates of the sites in the AWACA transect.

	D17	D47	D85	DMC
Latitude [°S]	66.72	67.39	70.43	75.10
Longitude [°E]	139.71	138.70	134.11	123.33
Altitude [m a.s.l.]	463	1530	2630	3233

2.2 Auxiliary data

In addition to radar data, this study uses the radio soundings at DDU and Concordia. At DDU, radiosondes are launched at 00:00 UTC every day by Météo-France. At Concordia, radiosondes are launched at 12:00 UTC every day and are coordinated by the Italian Antarctic Meteo-Climatological Observatory (IAMCO) (Grigioni et al., 2022). ERA5 data (Hersbach et al., 2020) at 0.25° and hourly resolution is used for contextual synoptic information over and around the transect. Finally, ice crystal images from the ICE-CAMERA (Del Guasta, 2022) were used to assist with the micro-physical interpretations at Concordia. The ICE-CAMERA takes the form of a flatbed scanner with a horizontal glass surface to collect precipitating particles. Images are taken with a resolution of 7 μm every 10 minutes, after which the plate is heated to sublimate the crystals.

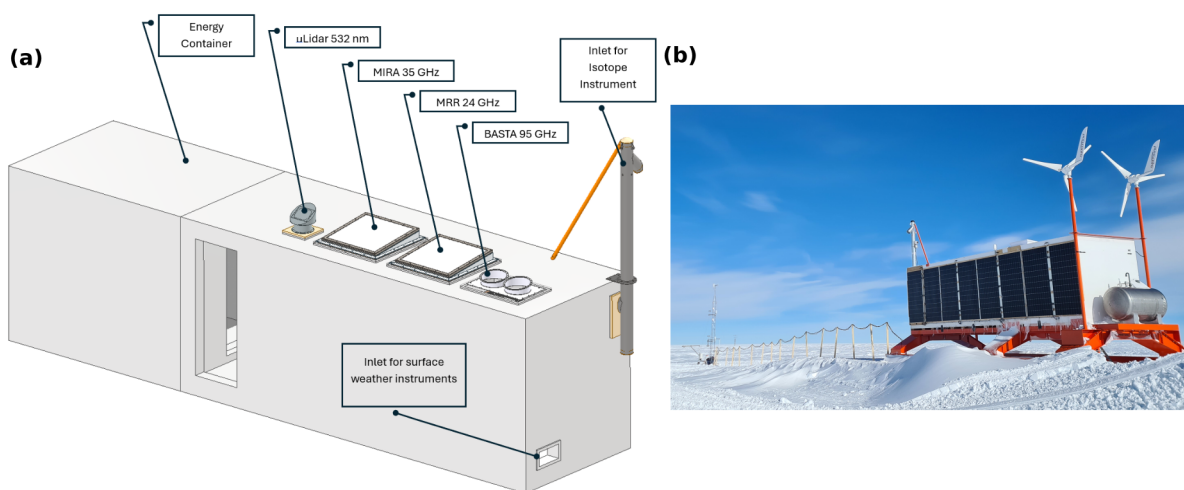


Figure 2. The Operational Platform Unit (OPU) of the AWACA transect. (a) shows the configuration at the energy-autonomous sites (D17, D47 and D85). The OPU at DMC has no energy container as the OPU is connected to the electrical network of Concordia station. (b) shows the OPU at D47. A weather station is positioned 30 m to the East of each OPU.

Table 2. Parameters of the radars in the AWACA OPUs. The BASTA has three modes with different range resolutions. The modes alternate every 3 s, giving a time resolution for each mode of 9 s.

	MRR-Pro	MIRA	BASTA
Transmission	FMCW	Pulsed	FMCW
Frequency [GHz] (band)	24.23 (K)	35.1 (K_a)	94.95 (W)
Time resolution [s]	5	3	9
Range [m]	0-6375	72-11512	40-12000
Range resolution [m]	25	24	12.5, 25, 100
Nyquist velocity [$m s^{-1}$]	0-5.85	± 13.2	± 9.9
Spectral resolution [$m s^{-1}$]	0.377	0.052	0.039
Dual-polarisation	no	yes	no

100 2.3 Radar data processing and products

This study principally makes use of the MIRA radar moments (Z_e , MDV, Doppler spectral width, LDR) in their original resolution (see Table 2). The highly sensitive MIRA is treated as a reference instrument for the inter-calibration of the three radars at each site. A discussion of the inter-calibration of the radars between the different sites is given in Appendix A4. To



remove a small amount of ‘speckle’ above the meteorological signal, the MIRA moments are filtered to regions for which the signal to noise ratio (SNR) is greater than -24 dB.

The BASTA and MRR are principally used to calculate the dual frequency ratios DFR_{K_a-W} and DFR_{K-K_a} respectively. For this calculation, the reflectivities are linearly interpolated to a common grid of 5 s and 25 m. The 25 m resolution mode of the BASTA is used to reduce the interpolation required. The reflectivities are corrected for gas attenuation using PAMTRA (Mech et al., 2020), with the required temperature, pressure and humidity input from ERA5. The radars at each site are then inter-calibrated using the calibration transfer methodology of Jorquera et al. (2023), resulting in the addition of a reflectivity offset to the BASTA and MRR data. At the coastal site D17, an additional offset is added to the BASTA reflectivity after an estimation of the attenuation due to liquid hydrometeors using the method of Dias Neto et al. (2019). The calibration and attenuation correction are described in Appendix A. The inter-calibrated and attenuation-corrected reflectivities are subtracted in logarithmic units to give the DFR.

The same corrections are applied to the spectral reflectivity data for the calculation of the spectral DFR_{K_a-W} . For this, the MIRA and BASTA spectra are linearly interpolated to a common grid of 25 m and 0.05 m s $^{-1}$. In addition to the reflectivity calibration, a range-dependent velocity correction is applied to the BASTA spectra to correct for the misalignment of the radar beam from vertical. If a radar does not point perfectly vertically, the measured Doppler-velocity will include a contribution from the horizontal wind. This contribution was found to be significant when calculating the spectral DFR, since a small misalignment can lead to a velocity offset of the peaks in the two spectra. The ‘velocity alignment’ used to calculate the offset is discussed in the article supplement.

To help identify regions of multi-modality in the Doppler spectra, the peakTree algorithm (Radenz et al., 2019) is applied to the MIRA data. Finally, an estimation of the rime mass fraction is calculated from the MIRA MDV according to the relation presented by Kneifel and Moisseev (2020). For this, the MDV is averaged to 20 minute and 100 m windows to reduce the effect of vertical air motions and adjusted to mean sea level pressure using pressure data from ERA5.

3 Overview of the case study

3.1 Synoptic Overview

The case study focuses on a moist intrusion resulting from a coastal cyclone which approaches DDU on the 15th of February 2025 (Fig. 3). Cyclonic flow on the eastern side of the low pressure center leads to the meridional transport of moist air parcels from the ocean onto the Antarctic plateau and along the observational transect. The moisture filament associated with the cyclone is classified as an atmospheric river (AR) according to the DARK framework of Buffet et al. (2026) but not according to the method of Wille et al. (2025) (V.Buffet, personal correspondence). However, the moisture transport close to the coast is not sufficient for the AR status to be retained as the cyclone approaches the continent. This event nevertheless corresponds to a strong advection of moist and warm air over the Antarctic ice sheet. At around 12:00 UTC on the 16th of February, the cyclone becomes occluded and starts to decay. This study focuses on the period between the arrival of the cyclone-driven cloud

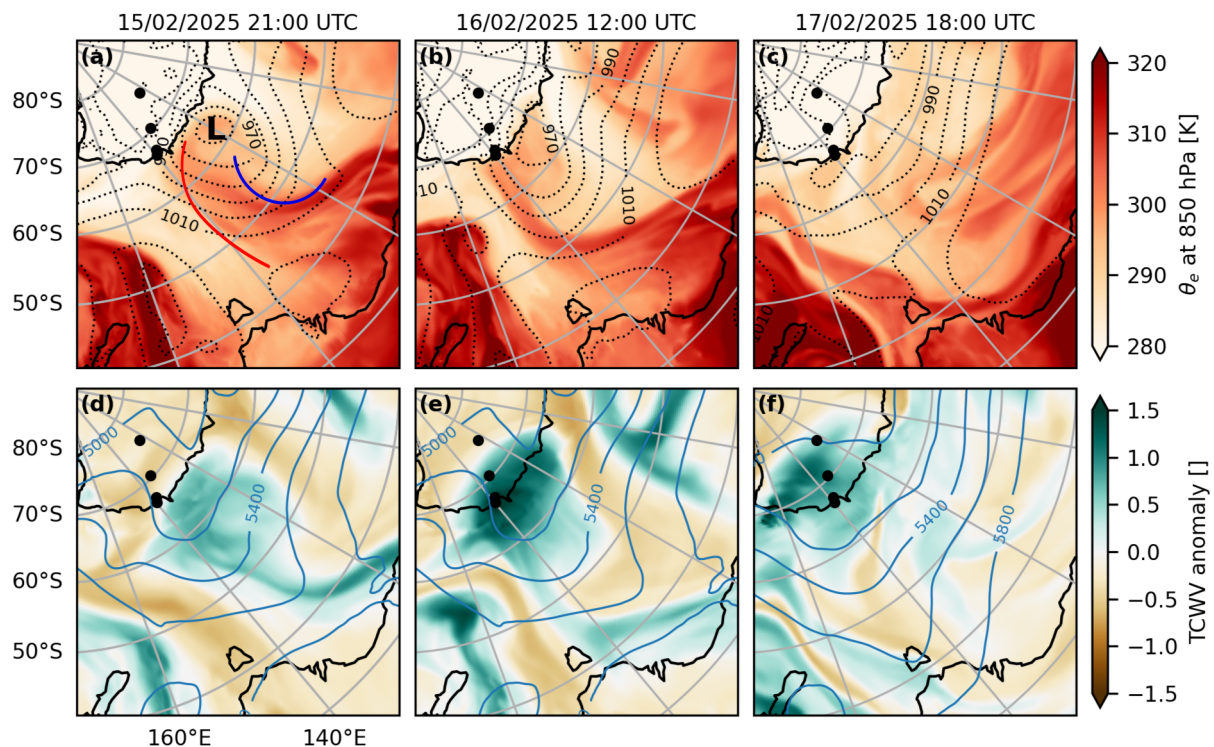


Figure 3. Synoptic maps for the case study. All data is from ERA5. The AWACA sites are shown as black dots. Panels (a)-(c) show the equivalent potential temperature θ_e at 850 hPa and mean sea level pressure (contours, 20 hPa intervals). The position of the low pressure center (L) and the warm (red) and cold (blue) fronts were added manually. Panels (d-f) show the total column water vapour (TCWV) plotted as an anomaly relative to the February 2025 mean $(TCWV - TCWV_{02/25})/TCWV_{02/25}$ in order to highlight the intrusion onto the plateau, where the absolute TCWV is low. Blue contours show the geopotential height in meters at 500 hPa.

and precipitation at the coastal sites (12:00 UTC on the 15th of February 2025) and the end of the 17th of February 2025, by which time the cyclone has decayed.

The transport of moisture inland requires the ascent of coastal air masses onto the Antarctic plateau. The origin of this ascent was investigated using Lagrangian trajectory analysis. The discussion below is based on a set of 24-hour forward and backward trajectories initiated at 20 hPa and 10 km intervals in a 100 km radius circle around D17 at 18:00 UTC on the 16th, using the LAGRANTO tool (Wernli, 1997; Sprenger and Wernli, 2015) and ERA5 data.

A first question is the degree to which the ascent onto the plateau can be attributed to synoptic ascent within the cyclone, rather than to ascent forced by the coastal slope. In mid-latitude cyclones, the principal ascending air stream is known as the warm conveyor belt (WCB), a coherent air stream which ascends diabatically from the boundary layer to the upper troposphere (Browning, 1986; Wernli, 1997). Studies of WCBs using Lagrangian trajectories typically use a threshold of 600 hPa of ascent in 48 hours (eg. Eckhardt et al., 2004; Heitmann et al., 2024; Gehring et al., 2020). This threshold effectively rules out

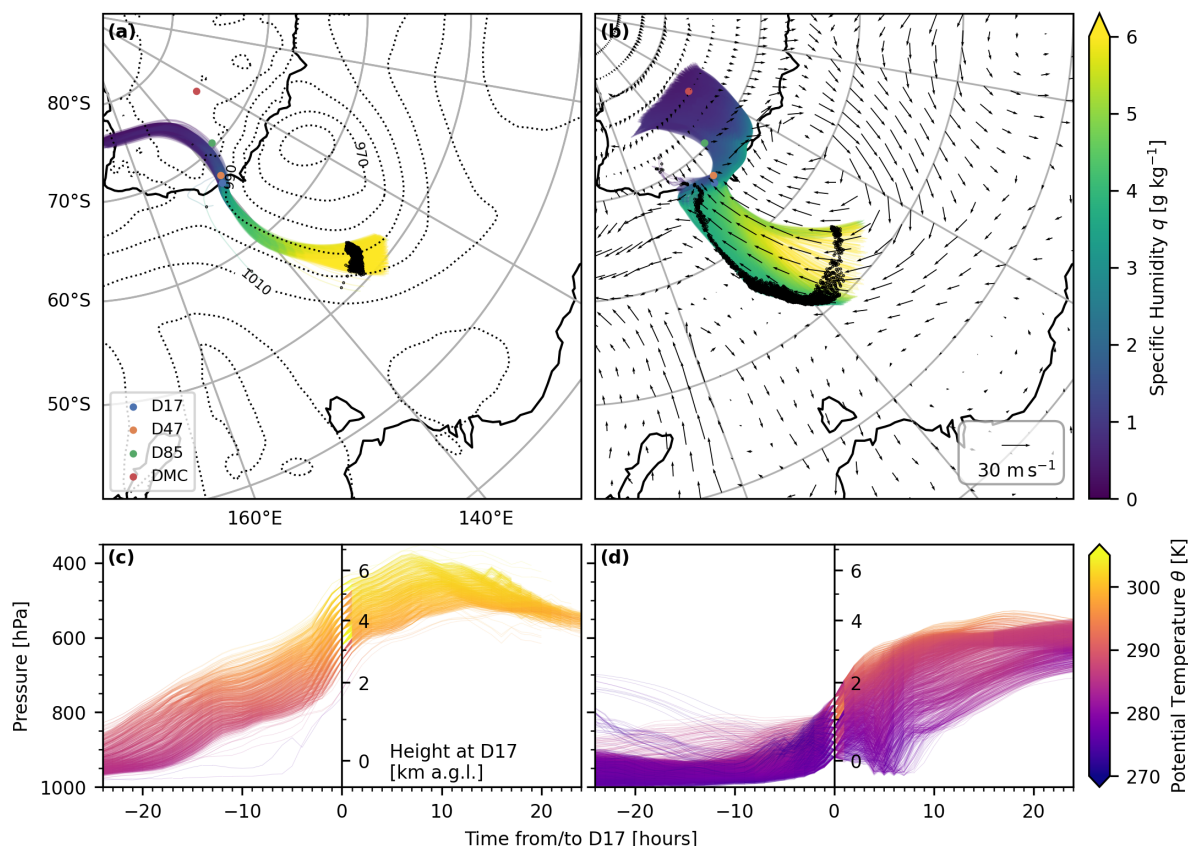


Figure 4. Lagrangian trajectory analysis of the case study using the LAGRANTO tool (Sprenger and Wernli, 2015) and ERA5 data. 24-hour forward and backward trajectories are started at D17 at 18:00 UTC on the 16th of February 2025. (a) and (c) show the trajectories which ascend more than 450 hPa in 48 hours. (b) and (d) show the trajectories which have a pressure of 750 hPa or greater at D17. In addition, (a) shows the mean sea level pressure and (b) shows the wind at 850 hPa, both at 21:00 UTC on the 15th of February 2025. The position of the trajectories at this time is shown with black dots.

WCBs associated with cyclones around Antarctica: the low ocean temperatures means that the quantity of water evaporated is insufficient to fuel strong ascent (Eckhardt et al., 2004; Madonna et al., 2014). However, although they do not appear on WCB climatologies, cyclones around Antarctica still contain an ascending stream of warm moist air. With this in mind, Fig. 4 shows the trajectories ascending more than 450 hPa in 48 hours. Although none of these trajectories ascend 600 hPa in 48 hours, it can be seen that the ascent starts before the coast and originates in the warm sector of the cyclone. This suggests that the ascent onto the plateau can, at least in part, be attributed to dynamical forcing linked to the warm air stream of the cyclone and not entirely to orographic forcing.



However, orographic forcing is still expected to contribute to the ascent onto the plateau. To investigate this, trajectories
155 arriving at D17 closer to the ground (at pressures greater than 750 hPa) are plotted (Fig. 4 (b) and (d)). These trajectories show
sharp ascent close to the coast, suggesting a strong orographic component.

The question remains as to how this orographic ascent is forced. The non-dimensional mountain height $\hat{H}$ indicates how
horizontal flow behaves upon meeting a tall obstacle (eg. Schär and Durran, 1997). If $\hat{H}$ is less than 1, the approaching flow
is unblocked and can proceed over the ‘obstacle’ of the coastal slope. Figure 4 (b) shows the wind speed at 850 hPa. Taking a
160 wind speed U of 30 m s^{-1} in the ‘jet’ perpendicular to the coast at 21:00 UTC on the 15th (Fig. 4 (b)), an obstacle height H
of 500 m (corresponding to the initial slope up to D17) and a Brunt-Väisälä frequency N of 0.02 m s^{-1} (calculated from the
radiosonde at DDU at 00:00 UTC on the 16th) gives $\hat{H} = (N \cdot H)/U \approx 0.3$, indicating unblocked flow. Diabatic heating from
freezing and condensation may also help drive the ascent. The increase in potential temperature along the trajectories indicates
that some diabatic heating occurred during the ascent, but the relative significance of this forcing is not known.

165 In summary, a coastal cyclone on the 15th of February 2025 led to moisture transport along the observational transect and
ascending air masses. The ascent can be attributed both to synoptic flow within the cyclone and to forced orographic ascent of
air masses incident on the coastal slope. Although the event does not meet the common criteria to be classified as a WCB or
AR when it reaches the coast, it still constitutes a significant intrusion of warm, moist air onto the high Antarctic plateau. The
alignment of air trajectories along the observational transect during the event allows the processes occurring at different stages
170 of the inland progression to be investigated and compared.

3.2 Macro-physical Precipitation Overview

This section presents the macro-scale features of the moist intrusion once it makes landfall, focusing on the radar data and
using ERA5 data for contextual information along the transect. Figure 5 shows the reflectivity, MDV and Doppler spectral
width measured by the MIRA radar at the four mainland sites of the transect from the 15th to the 17th of February. The onset
175 of clouds and precipitation at each site is offset as the intrusion moves along the transect. The time delays agree with those
calculated using LAGRANTO trajectory analysis: starting at 700 hPa at 18:00 UTC on the 16th, trajectories take approximately
1 hour to reach D47 and 5.5 hours to reach D85. The proximity of D17 and D47 results in a similar cloud time-height profile at
the two sites, with noticeable features (such as the lower cloud top height at 06:00 UTC on the 16th at D17) showing an offset
of up to 1 hour. Forward trajectories from D17 eventually veer to the East and do not reach DMC. Back-trajectories from DMC
180 (not shown) reveal that the precipitating air masses arrive from the area westward of the AWACA transect. However, they still
originate from the same coastal cyclone and must undergo a similar coast-plateau transition as those travelling over D17.

One of the clearest differences in radar data between the four sites is the decrease in reflectivity from the coast to the plateau.
Assuming a consistent calibration of the MIRA across the four sites (see section 2.1 and Appendix A4 for a discussion), the
decrease in reflectivity signals a decrease in the size and/or number concentration of particles as we move inland and hence in
185 a decrease in snowfall rate S . Although snowfall quantification is not the aim of this paper, a quick calculation using existing
 $Z_e - S$ power-law relations can be made. Table 3 shows the average snowfall rates for the first 24 hours after the arrival of
precipitation at each site, according to the relations of Matrosov (2007) and Falconi et al. (2018). The calculation was performed

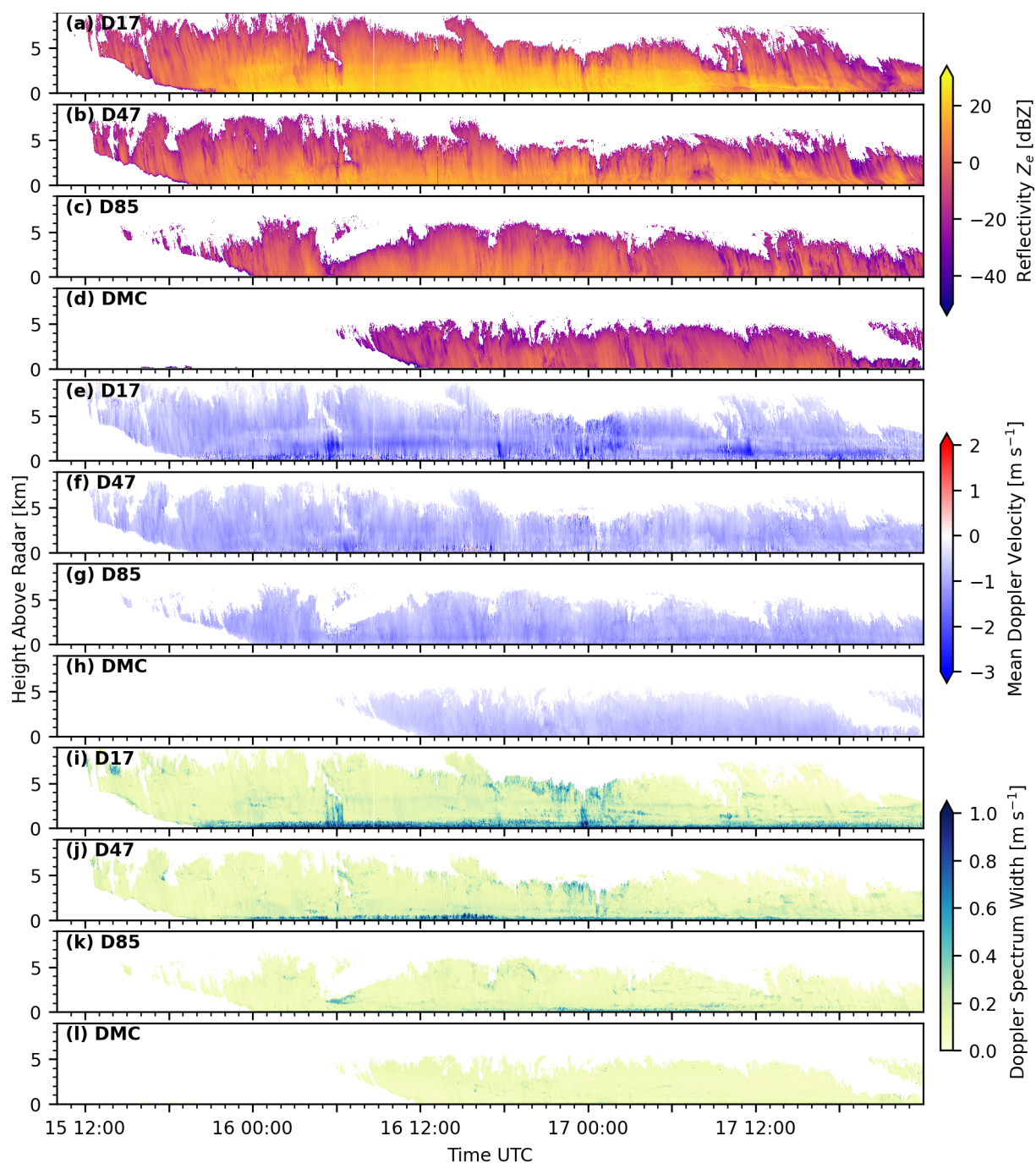


Figure 5. Reflectivity, mean Doppler velocity and spectral width measured by the MIRA radar at the four sites D17, D47, D85 and DMC during the case study.



Table 3. Snowfall rate estimates for the moist intrusion case study. The mean reflectivity Z_e and snowfall rate S (in mm of water equivalent per hour) were calculated for the 24 hours after the arrival of precipitation at each site, at one range above the turbulent layer. The parameters for the snowfall rate calculation were taken from Matrosov (2007) (M2007) and the low, medium and high riming-fraction parameters of Falconi et al. (2018) (F2018 LR, MR and HR).

	MIRA				BASTA			
	D17	D47	D85	DMC	D17	D47	D85	DMC
Reflectivity Z_e [dBZ]	21.5	12.0	5.1	-2.1	17.3	9.2	4.4	-0.9
S M2007 [mm hr ⁻¹]	2.1	0.3	0.1	0.02	8.6	0.8	0.2	0.05
S F2018 LR [mm hr ⁻¹]	3.2	0.4	0.1	0.02	6.2	0.9	0.3	0.08
S F2018 MR [mm hr ⁻¹]	5.2	0.4	0.1	0.01	12.8	1.2	0.3	0.06
S F2018 HR [mm hr ⁻¹]	7.3	0.4	0.0	0.01	15.3	1.2	0.3	0.05

using the reflectivity at one range above the near-surface turbulent layer (to avoid the potential impact of blowing snow) at each site: 1000 m, 800 m, 500 m and 150 m a.g.l. for D17, D47, D85 and DMC respectively. These ranges were obtained from the time-height plots of spectral width (Fig. 5). The results highlight the strong decrease in S inland along the transect, albeit with a large degree of variation according to the parameters and wavelength. However, it is well-documented that the parameters of $Z_e - S$ relations are site-specific and vary depending on the snowfall process and particle size and shape (eg. Souverijns et al., 2017; Scarchilli et al., 2020; Bracci et al., 2022), so the obtained values should be considered as orders of magnitude. The quantification of snowfall fluxes using data from the AWACA OPU's will be the subject of future work.

The MDV and spectral width reveal a second trend: near-surface atmospheric turbulence decreases in vertical extent as the intrusion moves inland. The height of the turbulent layer near the surface decreases from around 800 m at D17 to less than 300 m at D85, and is not evident at DMC. The tall turbulent layer at D17 is linked to the strong katabatic wind of the coastal slope, a common feature in the vicinity of DDU (eg. Wendler et al., 1997; Naithani et al., 2001; Grazioli et al., 2017b). The low-level turbulence at the coastal sites is consistent with the detection of blowing snow by the sensors on the meteorological towers at D17 and D47 (not shown). Turbulent features are also present at the cloud top and within the cloud at all sites, though with higher prevalence at the coastal sites. The identification of their origin is beyond the scope of the present study.

Cross sections of ERA5 data (Fig. 6) provide additional context on the atmospheric conditions along the transect during the intrusion. For example, the meridional wind v reveals that a katabatic layer is present on the coastal slope throughout the event, despite the general inland movement of air masses. The turbulent nature of the katabatic wind will contribute to the turbulent lower layers observed by the radars at D17 and D47. The cross section of vertical wind highlights the strong ascent of air masses over the coastal slope. This is a Eulerian snapshot of the ascent explored with Lagrangian trajectories in section 3.1. It shows that the ascent is constrained to the coastal slope and, though still present over D47, is significantly stronger over D17. A similar distribution is observed for the cloud liquid water content (CLWC). This similarity is not unexpected: supercooled liquid water is expected in regions of strong uplift, since the rapid ascent cools water vapour faster than ice nucleation can

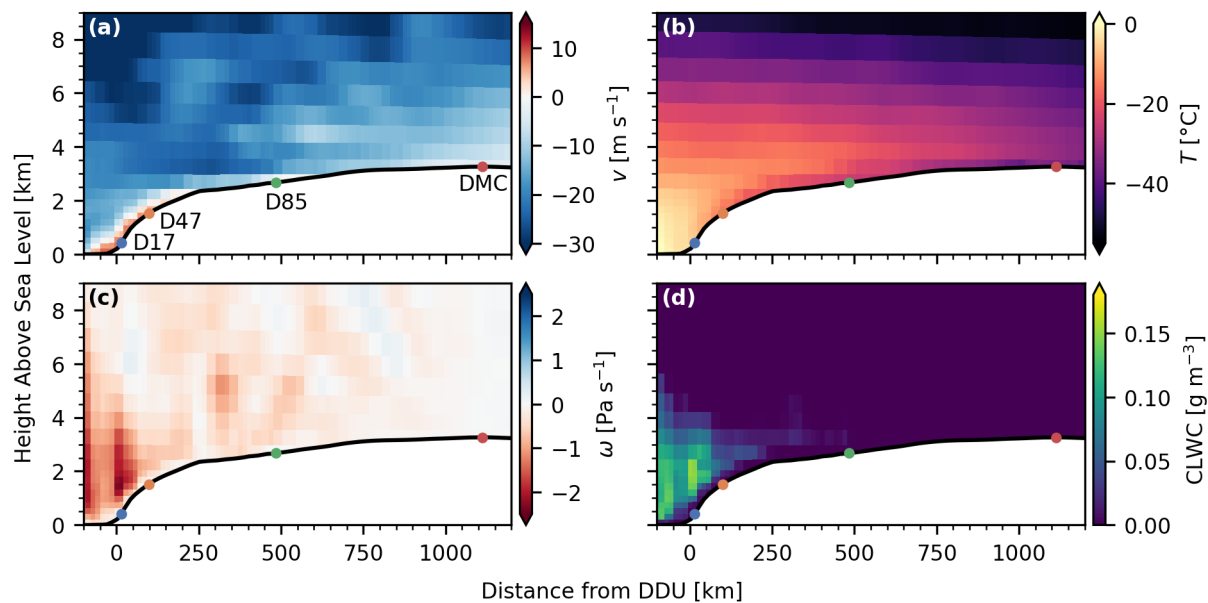


Figure 6. Vertical cross sections of (a) the meridional wind component v , (b) the temperature T , (c) the vertical wind ω and (d) the cloud liquid water content (CLWC) from ERA5 data along the AWACA transect. All four cross-sections show data at 12:00 UTC on the 16th of February 2025.

210 occur, particularly in an environment where ice nucleating particles are scarce (Sauerland et al., 2024; Wex et al., 2025). In addition, the higher temperatures at the coast and the proximity to the oceanic moisture source mean that the presence of supercooled liquid water is more likely over the coastal slope than over the high plateau. This is important as the presence of liquid water constrains the sites at which riming can occur. The low spatial and temporal resolution of ERA5 means that the variables must be interpreted with caution, particularly close to the coast where coarsely-resolved topography and coastline
215 lead to higher uncertainties in an area of strong local gradients. Despite this, it remains useful for providing a meteorological context for the radar observations and providing loose constraints on the possible micro-physical processes at the different sites.

In the next sections, we present the radar data site by site as the moist intrusion travels inland along the observational transect. For each site, we first focus on time periods which we believe are representative of the processes occurring throughout
220 the majority of the event. The peakTree algorithm (Radenz et al., 2019) was used to help identify these regions by highlighting periods of multi-modality in the MIRA spectral data. After discussing the first two sites (D17 and D47), we also zoom in to a period of strong multi-modality present at both of these sites (Section 6).

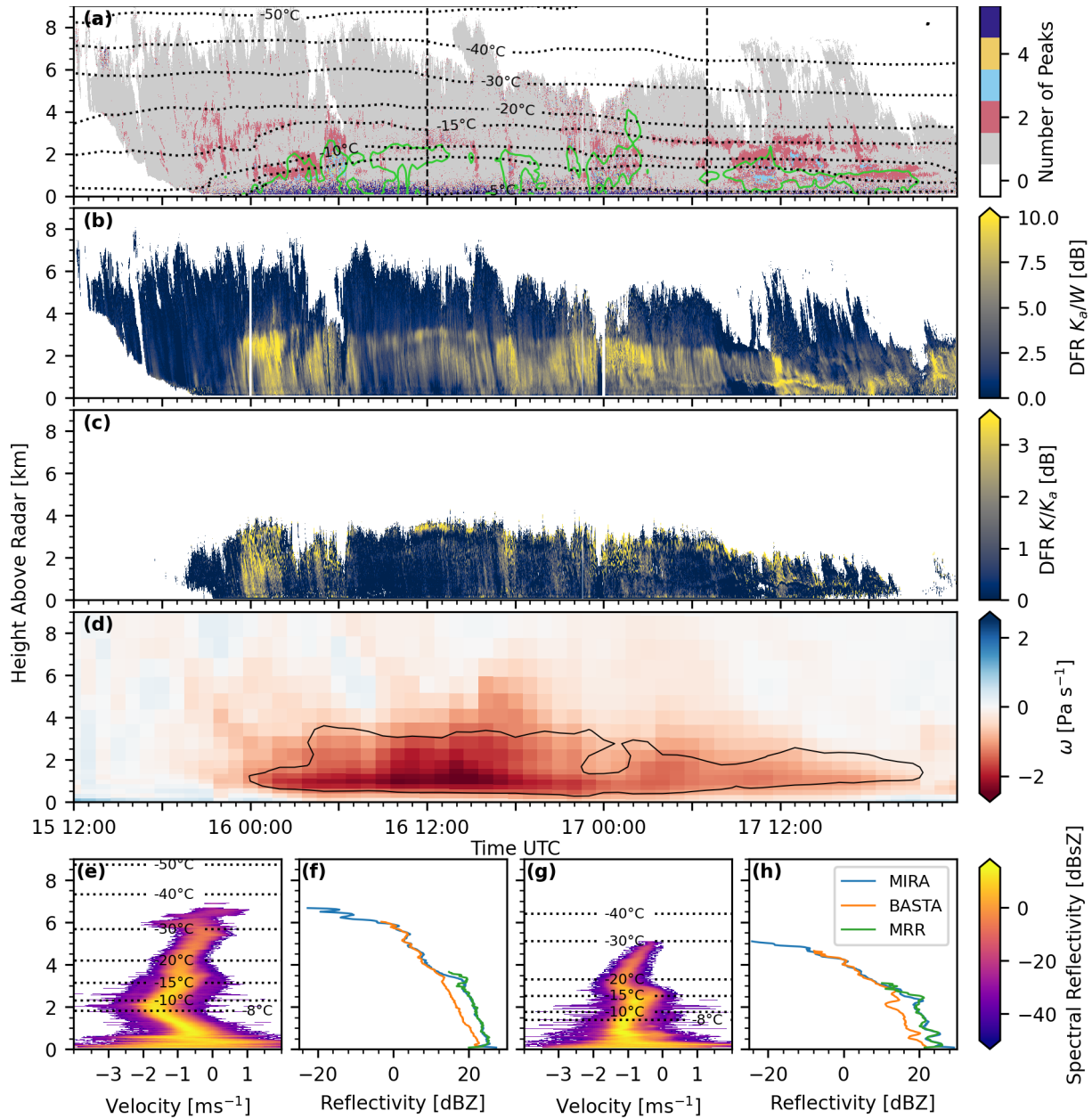


Figure 7. Radar data overview at D17 for the 15th-17th of February 2025. (a) The number of peaks in the Doppler spectrum calculated using peakTree. Black dotted contours show the temperature from ERA5. The 0.4 rime mass fraction contour is plotted in bright green (Kneifel and Moisseev, 2020). Dashed black lines indicate the times of the spectrograms and reflectivity profiles: 12:00 on the 16th for (e) and (f) and 07:00 on the 17th for (g) and (h). (b) Dual Frequency Ratio DFR between the MIRA and BASTA. (c) DFR between the MRR and MIRA. (d) Vertical velocity and cloud liquid water content (black contour at 0.05 g m^{-3}) from ERA5.



4 D17: 10 km from the coast and 463 m a.s.l.

We first focus on the coastal site D17 and discuss the dominant features representative of the whole event. To complement the
225 reflectivity and Doppler velocity information in Fig. 5, Fig. 7 shows the number of peaks detected in the Doppler spectrum by
the peakTree algorithm, the two DFRs and selected spectrograms and reflectivity profiles. For context, we also show the ERA5
vertical velocity and CLWC plotted in the same time-height space as the radar data.

During the three-days, several persistent features stand out. Firstly, the region of consistently high reflectivity after precip-
itation reaches the ground, which extends up to 3.5 km a.g.l. on the 16th and 3 km a.g.l. on the 17th. The high reflectivity
230 corresponds to a region of high DFR_{K_a-W} , indicating large particles outside the Rayleigh regime in the W-band. The top of
this region corresponds to an apparent ‘band’ of reduced MDV (Fig. 5) between 2 and 3.5 km a.g.l. on the 16th, and between
2 and 3 km a.g.l. on the 17th. A second region of lower MDV is present at lower ranges. The fact that the MDV decreases in
these regions despite the large particle size implied by the reflectivity and DFR_{K_a-W} indicates that updrafts affect the particles
in the radar beam. However, comparison with the ERA5 vertical velocity shows that rather than being constrained to distinct
235 bands in altitude, the updraft region instead extends uninterrupted from the surface. This highlights the dual influences on the
measured MDV: the terminal fall speed of the scattering particles within the cloud, and the vertical air motion. The resulting
‘zig-zag’ in altitude of the MDV is particularly evident in the spectrogram at 12:00 on the 16th (Fig. 7 (e)). The increase in
particle fall speed initially outweighs the updraft, leading to an increasing (increasingly negative) MDV between 3.5 and 2 km
a.g.l. At lower ranges, the updraft dominates and the measured MDV decreases.

240 Two questions arise from this velocity structure: the origin of the updraft and the micro-physical processes which lead to
a sufficiently large increase in terminal fall speed that the MDV increases despite upward air velocities. For the former, the
origin of the updraft on the coastal slope was explored in section 3.1. It can be attributed to a combination of large-scale
uplift due to the coastal cyclone and uplift forced by the orography of the coastal slope. For the micro-physical process, the
high terminal fall speed points to the occurrence of riming: the freezing of supercooled liquid particles onto ice and snow
245 particles causes an increase in particle density and hence fall speed (eg. Barthazy and Schefold, 2006). The presence of the
supercooled liquid water required for riming is consistent with the ERA5 data (Fig. 6 and Fig. 7). In addition, the observation
of additional attenuation in the BASTA data after the onset of precipitation is indicative of the presence of liquid hydrometeors
(see Appendix A3). Applying the relation between Doppler velocity and rime mass fraction derived by Kneifel and Moisseev
(2020) to the MIRA MDV indicates that the MDV is consistent with elevated rime mass fractions as high as 0.8. Although
250 the quantitative values derived from this relation should be interpreted with caution, the consistently high values can be used
qualitatively to support the hypothesis of riming being an important process at D17 throughout the case study.

The DFR signatures associated with the region of updraft at D17 warrant further investigation. On the first order, the increase
of DFR_{K_a-W} at the top of the updraft region is consistent with the presence of large particles. However, this is not the only
signal visible in the time-height plots. Firstly, neither DFR_{K_a-W} nor DFR_{K-K_a} show a strong increase *within* the updraft
255 region, but rather at the top of it. This could indicate either a lack of further particle growth within the updraft region, or
alternatively a saturation of the DFR signal due to the growth of very large particles. In addition, for much of the 16th, the



DFR_{K-K_a} appears to increase *above* the increase in DFR_{K_a-W}, before both DFRs stabilise or decrease. This is also visible in the reflectivity profiles, where Z_{eK} increases at a higher altitude than Z_{eK_a} . This is not expected for particles growing uniformly from above, since the Rayleigh regime extends to larger particle sizes at larger wavelengths. To investigate this signal, scattering cross-sections for W, K_a and K band were simulated as a function of particle size for several representative ice particles using PAMTRA (Mech et al., 2020). The results (Appendix B) show that the only way that DFR_{K-K_a} can increase at a smaller particle size than DFR_{K_a-W} is if the particles start with a diameter of around 4 mm. At such particle sizes, further particle growth can lead to a Mie resonance peak in DFR_{K-K_a} before DFR_{K_a-W}. This suggests that large particles are entering the radar beam without uniform growth from the top of the cloud. It could be explained by the advection of large particles into the radar beam, consistent with a region of strong vertical and horizontal air velocities.

The MIRA spectrograms during the period of updraft (Fig. 7) reinforce this picture of a dynamic, turbulent structure. In addition to the strong inflection in the mean velocity, the spectrograms also show substantial turbulent broadening throughout the lowest 3 km. Despite this broadening, a distinct secondary peak can often be identified towards the top of the updraft region (around 2-3 km a.g.l.). The low fall speed and low spectral LDR of this peak suggest that it may correspond to supercooled liquid water droplets, which would be consistent with the occurrence of riming. The peakTree analysis shows that the spectrograms presented have a modal structure which is representative for much of the 16th. In addition, the occurrence of this secondary peak is likely to be underestimated, since without a distinct minimum between the two peaks it will not be detected by the peakTree algorithm. Several periods of higher multi-modality are detected by peakTree. For example, from 05:00 to 06:30 on the 16th. These periods correspond to a high MDV and hence a higher estimated rime mass fraction. They will be explored alongside the equivalent time periods at D47 in section 6

5 D47: 94 km from the coast and 1530 m a.s.l.

Figure 8 shows the equivalent event-wide view for D47, with spectrograms chosen to be representative for the majority of the event. This section focuses on the similarities and differences compared to D17 on the event scale. Firstly, although ERA5 indicates upwards air velocities at D47, the magnitude is lower than at D17. As a result, the MDV and spectrograms do not show the characteristic ‘zig-zag’ pattern seen at D17. Similarly, ERA5 suggests liquid water is present but in smaller quantities than at D17. We therefore expect the role of riming to be less important for particle growth at D47. This is supported by the estimated rime mass fraction, which remains below 0.4 apart from an enhanced ‘pocket’ between 06:00 and 07:00 UTC on the 16th. The DFR_{K_a-W} increases below 3 km but not to the same extent as at D17, indicating generally smaller particle sizes. The DFR_{K-K_a} (not shown) remains low, indicating that particles are not exiting the Rayleigh regime at the K_a band, apart from within the same time period on the 16th. This time period will be examined in section 6.

A notable feature in the temperature structure is the presence of a quasi-isothermal region at -15°C in the 3 km next to the ground. A temperature of -15°C falls within the dendritic growth zone, within which ice growth from the vapour is favourable (eg. von Terzi et al., 2022). Dendritic growth would lead to oblate, irregular snowflake shapes with low densities, consistent with the generally low MDV observed by the MIRA (Fig. 5). It is therefore probable that dendritic growth is contributing to

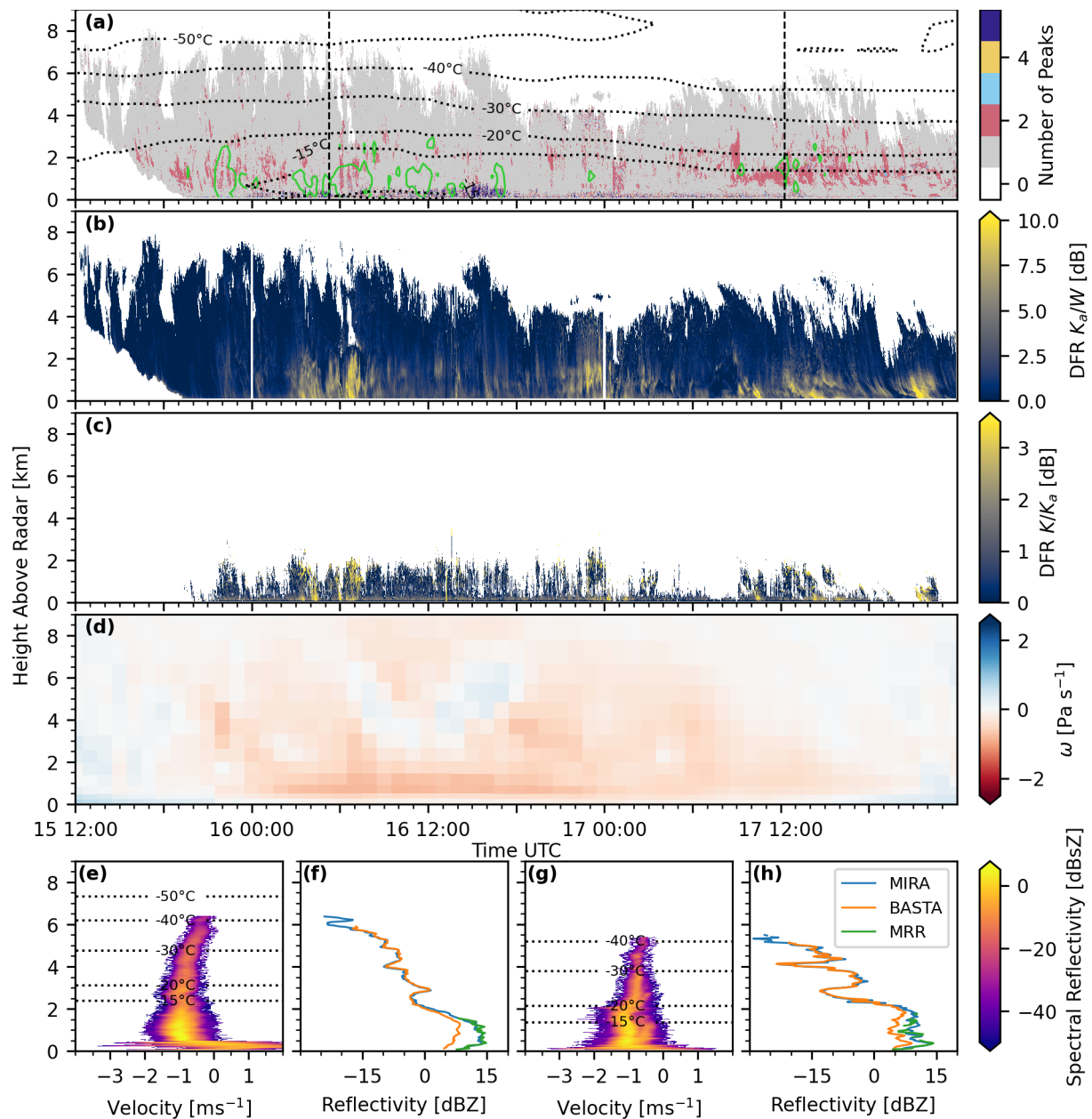


Figure 8. As in Fig. 7 for D47. In (a), the rime mass fraction contour at 0.2 is plotted in green contours (note a lower value than at D17). The spectrograms and reflectivity profiles are at 05:15 on the 16th (e and f), and 12:15 on the 17th (g and h).

290 particle growth at D47. Indeed, for much of the event the strongest increases in reflectivity and DFR_{K_a-W} occur at the top of this isothermal region. However, the isothermal region also corresponds to the region of updraft as indicated by ERA5. Hence it is unclear whether this increase is caused by enhanced riming within the updraft region, as hypothesised at D17 (albeit with a



lower availability of liquid water), or enhanced dendritic growth within a favourable temperature regime. In addition, dendritic particles have a high aggregation efficiency and aggregation could contribute to further particle growth within this layer.

295 The spectrograms at D47 generally show less turbulence-induced broadening than at D17 and a relatively low increase in MDV. However, an increase in terminal fall speed of the particles may be partially offset by the upward air velocities. In addition, a secondary peak is often visible (similar to the example at 12:15 on the 17th in Fig. 8(g)). The non-negligible reflectivity and velocity of this peak suggests it corresponds to ice fragments rather than small supercooled water droplets. The origin of the peak is unclear. However, it could correspond to increased variation in the ice particle distribution arising due to
300 dendritic growth and aggregation, including possible SIP from collisions between larger particles. A similar peak is observed in the dendritic growth zone by both von Terzi et al. (2022) and Teisseire et al. (2025). It may also correspond to primary ice production as a result of enhanced supersaturation in the updraft region.

6 Strong Riming Periods at D17 and D47

As identified in the discussion above, regions of higher MDV and hence estimated rime mass fraction are present at both D17
305 and D47. In particular, from 05:00 to 06:30 on the 16th at D17 and shortly afterwards (06:00 to 07:00) at D47. Further periods of enhanced rime mass fraction are observed at D17 (17:00 to 18:00 on the 16th, 23:00 on the 16th to 02:00 on the 17th, 11:00 to 13:00 on the 17th) but these are less evident at D47. This section concentrates on the first of these periods, zooming in to the polarimetric and spectral signatures to investigate the similarities and differences compared to the rest of the event.

Figure 9 shows Z_e , MDV, LDR, DFR_{K_a-W} , DFR_{K-K_a} and the number of Doppler spectrum peaks for 04:30 to 07:00 on the
310 16th at D17. The period from 05:00 to 06:30 shows higher reflectivity and velocity than the periods directly before and after, implying larger and faster-falling particles. The strong increase in particle size is confirmed by DFR_{K_a-W} , which increases up to 10 dB, indicating particle sizes outside the Rayleigh regime for W-band. DFR_{K-K_a} is highest near the top of the MRR sensitivity range but decreases below. This could signal that particles also exit the Rayleigh regime at K_a band, leading to a decrease in DFR_{K-K_a} with further particle growth, as discussed in Section 4. The LDR is particularly low from 05:15 to 06:30,
315 consistent with the more spherical form of heavily rimed particles.

The spectrogram at 05:30 (Fig. 10) at D17 shows three modes, the fastest of which shows a strong increase in fall-speed towards the ground. Such a strong acceleration could be attributed to a dynamical downdraft. However, the fact that the slowest falling mode remains close to 0 m s^{-1} over all ranges suggests that this acceleration is not only dynamical – a downdraft would cause all particle populations to increase in fall-speed – but rather corresponds to a micro-physical effect causing the
320 faster-falling population to increase in fall speed. The spectrally-resolved LDR does not show a very strong distinction between the different modes but is relatively low for all modes, consistent with more spherical shapes. Finally, the spectral DFR_{K_a-W} confirms the large difference in particle size between the fastest falling mode and the secondary particle population.

The spectral signatures are consistent with the riming of snowflakes by supercooled water droplets, resulting in a rapid increase in the density and fall speed of the largest particles. The mode which remains close to 0 m s^{-1} is consistent with
325 small supercooled liquid water droplets which would drive the riming. The third mode which appears below 2 km could signal

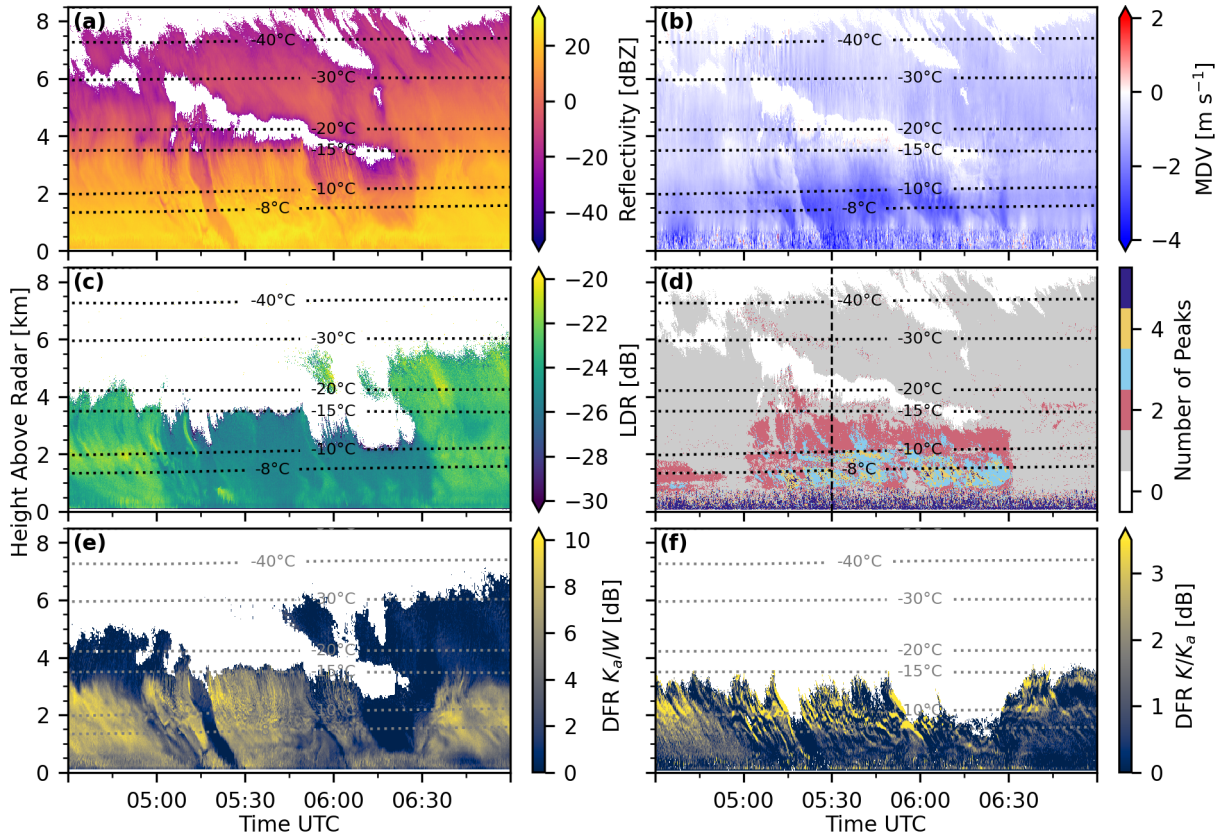


Figure 9. Radar data at D17 for the period 04:30-07:00 on the 16th of February 2025. Temperature contours are taken from ERA5. The vertical black dashed line on panel (d) shows the time of the spectrogram in Fig. 10. Temperature contours are from ERA5.

a particle population arising from SIP during the riming process. The temperature of -3°C to -8°C is consistent with the seminal laboratory study of secondary ice production via the Hallett–Mossop (rime-splintering) process (Hallett and Mossop, 1974), whereby ice fragments are released as supercooled droplets accrete onto the riming particle. However, it should be noted that a more recent study found no evidence for efficient rime-splintering in this temperature range (Seidel et al., 2024).

330 The general structure of two modes above 2 km and three modes between 1 km and 2 km above the ground is confirmed by peakTree to be persistent for much of the period from 05:00 to 06:30. This is in contrast to the rest of the event, for which typically one or two modes are present (Fig. 7). In addition, the spectrogram at 05:30 does not show the ‘zig-zag’ in velocity seen in other spectrograms, but rather a uniform increase in fall speed of the fast-falling mode. This suggests that the updraft is temporarily weaker, although this is not evident in the ERA5 data.

335 Figure 11 shows the equivalent time period at D47 (offset by one hour from D17). Similar to D17, there is a patch of enhanced Z_e , MDV and DFR_{K_a-W} from 06:30 to 07:30, consistent with large and fast-falling particles. The DFR_{K-K_a} shows some reaction during this time period, whereas it is very low for the rest of the event at D47 (not shown), suggesting that the

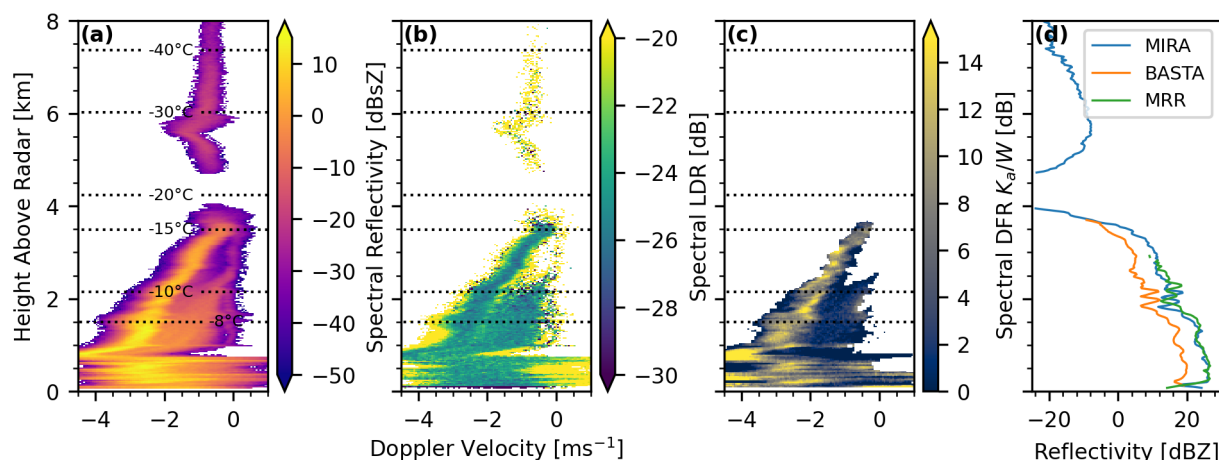


Figure 10. Spectrograms at D17 at 05:30 on the 16th of February 2025: (a) spectral reflectivity and (b) linear depolarisation ratio (LDR) from the MIRA. (c) shows the spectral dual frequency ratio between the MIRA and BASTA (DFR_{K_a-W}). Temperature contours are from ERA5.

particles start to exit the Rayleigh regime at K_a band. A strong reaction is also visible in the LDR, which is lower during the period 06:30 to 07:30 than in the periods before and after. The structure of LDR with height also changes: from a general increase close the ground before 06:30 to a decrease towards the ground after 06:30. A decrease in LDR towards the ground could signal particles becoming more spherical as they fall due to riming. The spectrogram at 06:45 (Fig. 12) shows strong similarity to that at 05:30 at D17 (Fig. 10). Notably, a strong acceleration of the fastest-falling mode and the presence of a secondary mode which remains close to 0 m s^{-1} in the lowest 2 km. As at D17, this is consistent with the riming of snowflakes by supercooled liquid water droplets.

This time period can therefore be interpreted as a ‘pocket’ of enhanced riming within the event at both sites. At D17, where riming is predicted to play a role throughout the event (see the rime mass fraction in Fig. 7), these periods may represent periods of increased riming intensity, arising for example from higher liquid water availability. At D47, where the estimated rime mass fraction is generally lower, this period may be one during which sufficient liquid water is available for riming to become the dominant process affecting particle growth. The environmental driver for the timing of these periods at D17 and D47 remains unclear. No strong changes are observed in the temperature, CLWC or updraft velocity in the ERA5 data. However, the hourly resolution may not be fine enough for these periods to be properly represented.

To summarise the discussion of the first two sites D17 and D47, the radar and temperature data are consistent with snow growth through riming and dendritic growth processes, with enhanced pockets of riming present at both sites. Despite the small horizontal distance between D17 and D47, the difference in position relative to the coastal slope results in differences in vertical temperature structure, turbulence and the availability of liquid water and hence differences in the dominant snow growth processes at the two sites.

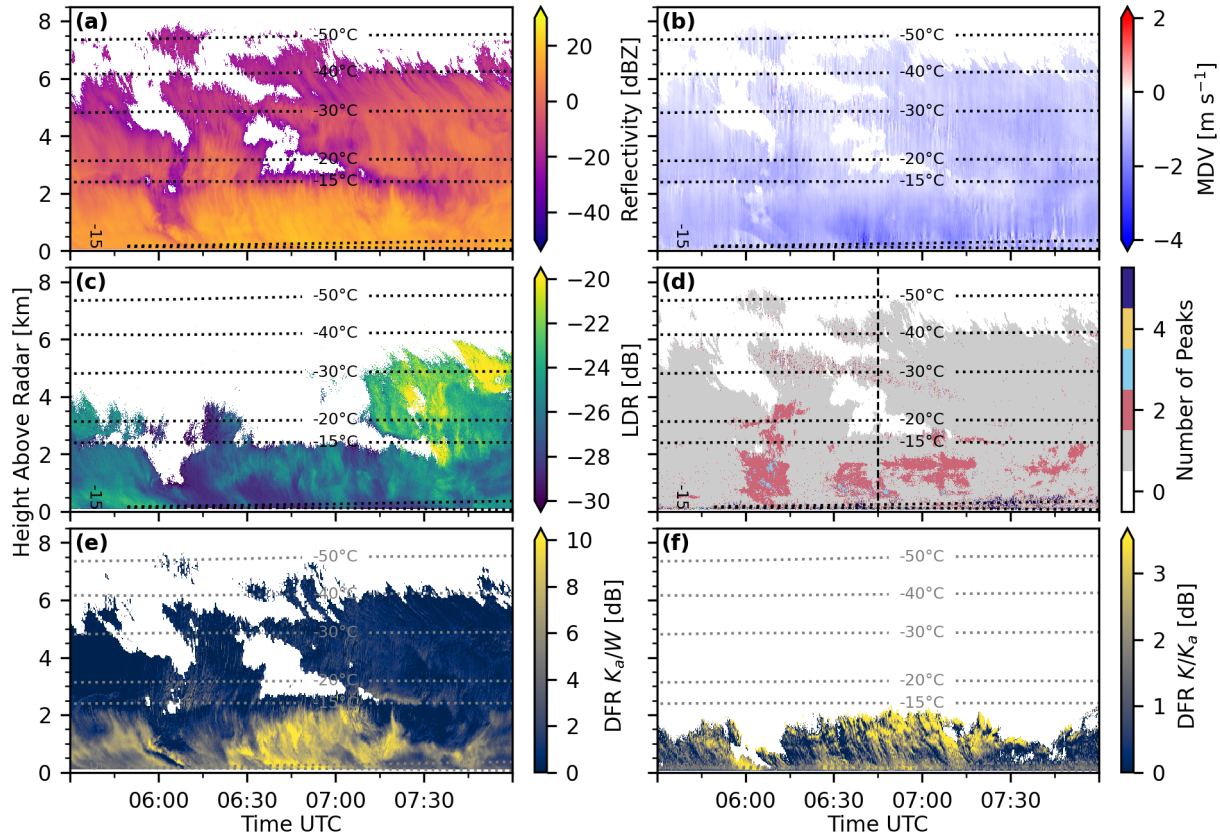


Figure 11. Radar data at D47 for the period 05:30–08:00 on the 16th of February 2025. Temperature contours are taken from ERA5. The vertical black dashed line on panel (d) shows the time of the spectrogram in Fig. 12.

7 D85: 477 km from the coast and 2630 m a.s.l.

The reflectivity (Fig. 5 and Table 3) at D85 remains significantly lower than at D17 or D47 for the duration of the event, signalling smaller particles. This is confirmed by DFR_{K_a-W} (Fig. 13), which remains below 2.5 dB throughout the event. However, some ‘fall-streaks’ of higher DFR_{K_a-W} are visible, indicating that some particles do exit the Rayleigh regime at W-band. The MDV (Fig. 5) and hence the calculated rime mass fraction (not shown) have a similar magnitude to that at D47 outside the riming pocket (≈ 0.2). However, the occurrence of riming is less likely at D85 due to the lower quantities of supercooled water. The ERA5 data (Fig. 6) suggests that liquid water does not extend up onto the plateau, with the maximum CLWC at D85 (0.008 g m^{-3}) being an order of magnitude below the typical values at D47. In addition, the rime mass fraction relation of Kneifel and Moisseev (2020) has a high uncertainty for MDVs below around 1.5 m s^{-1} .

A particularly striking feature at D85 is the presence of extended periods of very low (-30 dB) LDR, extending through the full height of the cloud (at least within the sensitivity of the cross-channel). For example, from 06:00 to 08:00 and 16:30 to 17:30 on the 16th and from 12:30 to 14:00 on the 17th. The first of these periods corresponds to a period with slightly

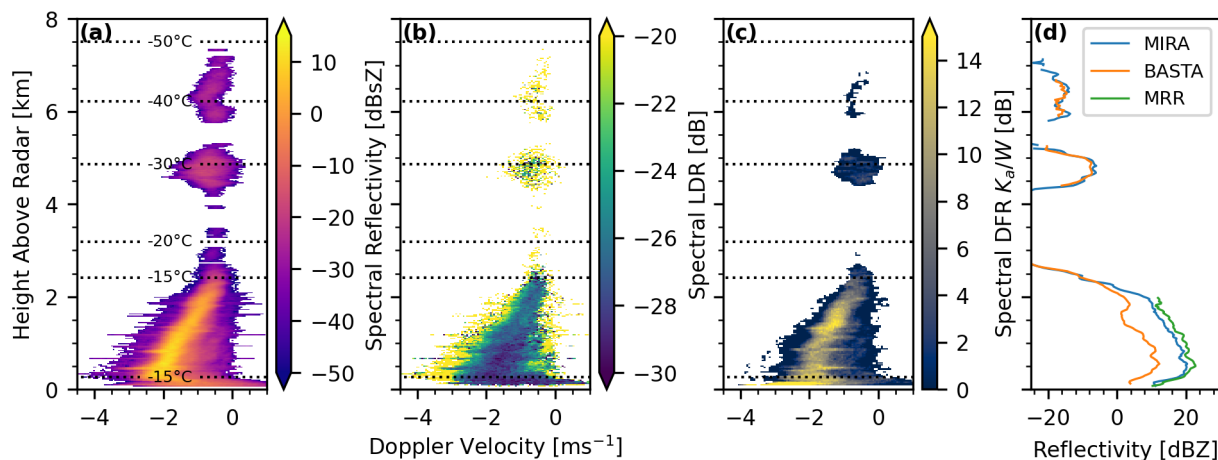


Figure 12. Spectrograms at D47 at 06:45 on the 16th of February 2025: (a) spectral reflectivity and (b) linear depolarisation ratio (LDR) from the MIRA. (c) shows the spectral dual frequency ratio between the MIRA and BASTA (DFR_{K_a-W}).

decreased reflectivity, whereas the other two periods do not show a strong reflectivity signature. In addition, no particular
 370 difference in MDV between the periods of low and high LDR is visible. This suggests that the change in LDR over time does
 not correspond to a strong change in particle size or terminal fall speed, suggesting that it is not linked to periods of strong
 particle growth through aggregation or riming. The ERA5 data (Fig. 13) suggests that the cloud-top temperature is between
 -40°C and -35°C , whereas the surface temperatures is between -25°C and -20°C . Hence the whole depth of the cloud
 lies within the temperature regime in which plate-like crystals are expected (Bailey and Hallett, 2009). One hypothesis for the
 375 periods of low LDR is that they correspond to periods with a higher proportion of horizontally-oriented single-plate crystals.
 This would be consistent with the simulations of Sato and Okamoto (2006), which show a low LDR for large plates compared
 to columns and bullet rosettes. The periods of higher LDR could contain a wider range of particle shapes, possibly including
 polycrystalline plate crystals or irregular aggregates. In addition, since the cloud top temperature lies close to the boundary
 between the column and plate regimes (around -40°C), the periods of higher LDR could also include crystals initiated with a
 380 columnar habit.

This hypothesis is partially supported by the spectrograms. During the periods of low LDR (eg. around 17:00 on the 16th,
 panels (d) and (e) in Fig. 13), the spectral LDR is lowest for the fastest-falling particles in a mono-modal spectrum, which
 could correspond to the largest plates in a particle population of relatively uniform habit. In the periods of higher LDR (eg.
 at 01:00 on the 17th, panels (f) and (g) in Fig. 13), there is a higher variation in the spectral LDR across the width of the
 385 spectrum and a greater prevalence of multi-modal spectrograms, suggesting a greater variation of particle shapes during these
 time periods, possibly including polycrystalline particles and small ice fragments. However, the hypothesis remains speculative
 due to the lack of in-situ particle images at D85. The question also arises as to what would cause strong temporal changes in
 the proportion of pristine plates. Small variations in the ambient temperature and humidity at the cloud top could lead to

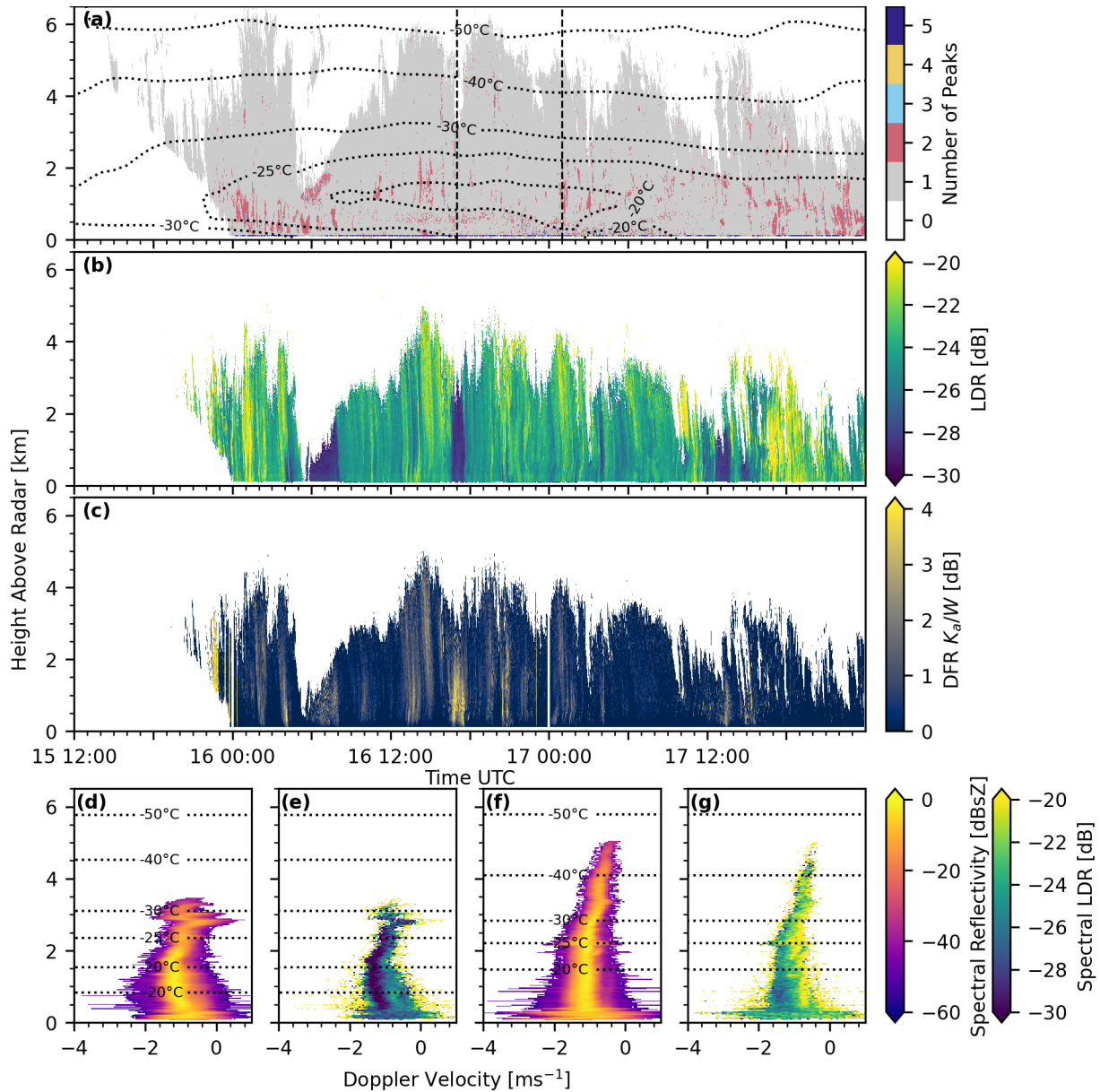


Figure 13. Radar data overview at D85 for the 15th-17th of February 2025. (a) The number of peaks in the Doppler spectrum calculated using peakTree on the MIRA data. Black dotted contours show the temperature from ERA5. Dashed black lines indicate the times of the spectrograms and reflectivity profiles: 17:00 on the 16th for (d) and (e) and 01:00 on the 17th for (f) and (g). (b) Linear depolarisation ratio (LDR) from the MIRA. (c) DFR between the MIRA and BASTA. For each of the two spectrogram times, the spectral reflectivity (d and f) and spectral LDR (e and g) from the MIRA are shown.

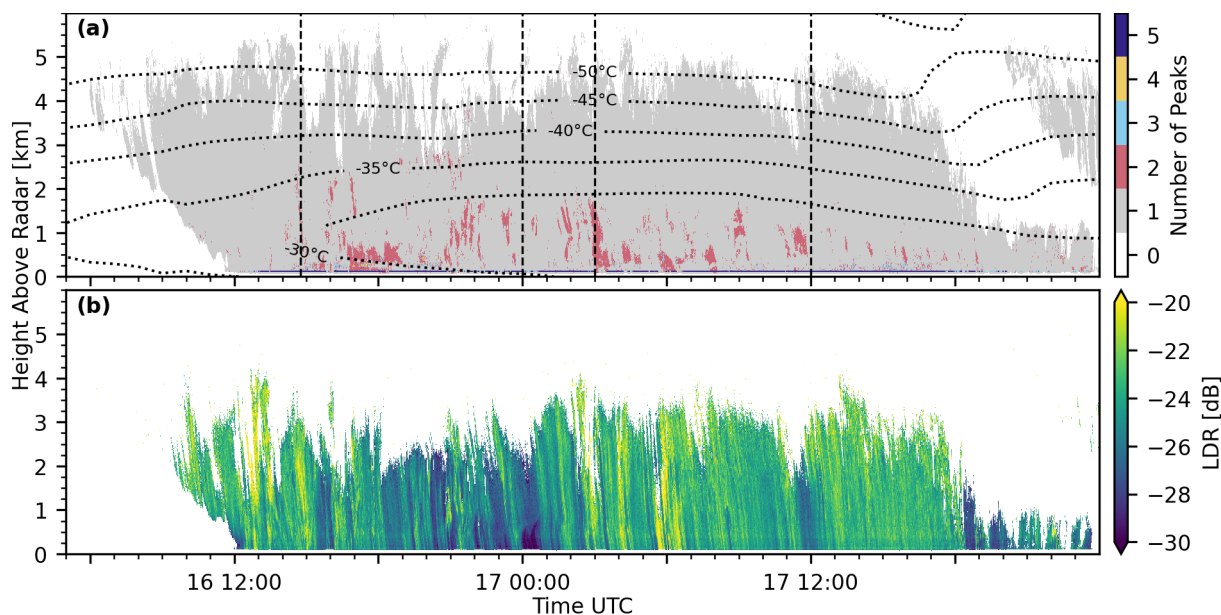


Figure 14. (a) The number of Doppler spectrum peaks identified by peakTree at DMC for the case study from the 15th to 17th of February 2025. Contours show the temperature from ERA5. Vertical dashed lines show the times of the four spectrograms in Fig. 16. (b) The Linear Depolarisation Ratio (LDR) from the MIRA radar.

variations in the particle habit, especially close to the ‘boundary’ between expected columnar and plate growth. However, no such thermodynamic changes are visible in the ERA5 data.

Returning to the secondary peak in the spectrograms (eg. Fig. 13(f)), the moderate fall velocity of this peak suggests that it corresponds to ice rather than supercooled liquid water. In fact, secondary peaks at D85 often resemble those observed at D47. However, the colder temperatures compared to D47 mean that only a small proportion of the cloud lies within the dendritic growth zone (the lowest 1.5 km is warmer than -20°C for the spectrogram in Fig. 13(f)). Although the secondary peak does appear to be constrained to this region, the small DFR signal suggests that dendritic growth is limited. Rather than dendritic fragments, we suggest that the multi-modality may arise from a mixed particle distribution of polycrystalline and pristine crystals (similar to at DMC, see Section 8), or new primary production in the lower part of the cloud.

In summary for D85, by the time the warm intrusion has penetrated almost 500 km into the Antarctic continent, the lower temperatures and lower moisture availability means that particles remain small, resulting in only small DFR signals in the radar data. Instead, interesting features arise in the LDR, which point to temporal evolutions in crystal growth habit. Despite the colder temperatures, the spectral structures show similarities to those at D47, implying that similar micro-physical processes are occurring. This points to D85 as a linking site between the coastal slope and the extreme cold of the high plateau.

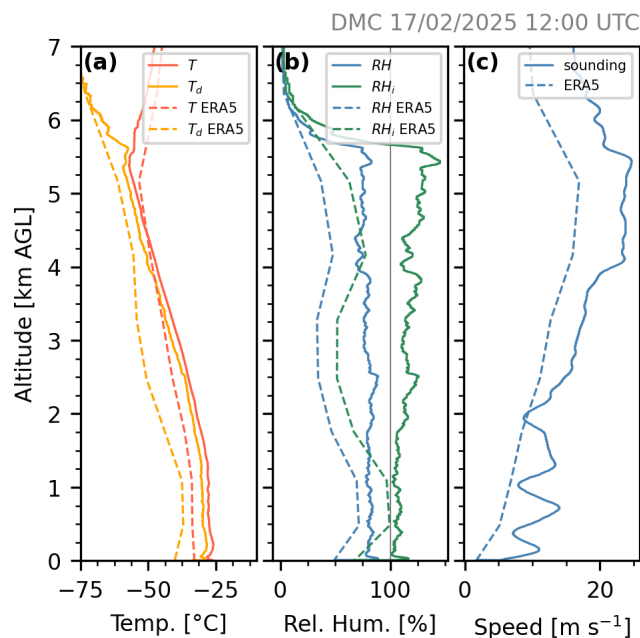


Figure 15. Radiosounding at Concordia station (DMC) at 12:00 UTC on the 17th of February 2025. ERA5 data is shown in dashed lines. (a) shows the Temperature T and dew-point temperature T_d . (b) shows the relative humidity with respect to liquid water (RH) and with respect to ice (RH_i). (c) shows the horizontal wind speed.

8 DMC: 1103 km from the coast and 3233 m a.s.l.

The reflectivity and MDV at DMC, the southernmost site of the transect, are smaller than at D85, continuing the pattern of a decrease in particle size along the transect. At DMC, there is no signal in the DFR_{K_a-W} (not shown), indicating that most particles are small enough to remain in the Rayleigh regime at W-band. We therefore expect that significant aggregation or riming processes are not occurring on the high plateau. Nevertheless, the radar data show interesting features. Figure 14 shows the LDR and multi-modality from the MIRA radar at DMC after the arrival of clouds and precipitation. Similar to D85, we notice strong changes in the LDR in time and some regions of multi-modality. At DMC, we have more contextual information to aid the interpretation of these features. Namely, the daily radiosonde and ice crystal images from the ICE-CAMERA. In the discussion below, we first use these two sources of information to find the expected and observed ice particle habits, then link these to the radar LDR and multi-modality signatures.

The radiosounding and ERA5 data (Fig. 15) show that the temperature increases from -55°C at the cloud top to -25°C at the surface. This range traverses the boundary between the growth regimes for columnar ($< -40^{\circ}\text{C}$) and plate-like ($> -40^{\circ}\text{C}$) ice crystals (according to Bailey and Hallett, 2009) at 3.5 km a.g.l. We can therefore expect columnar growth near the cloud top and plate-like growth close to the ground. However, the ICE-CAMERA images for this time period (Fig. 17) are dominated by the columnar regime: hollow columns, bullet-ended columns and polycrystalline bullet-rosettes, as well as numerous small

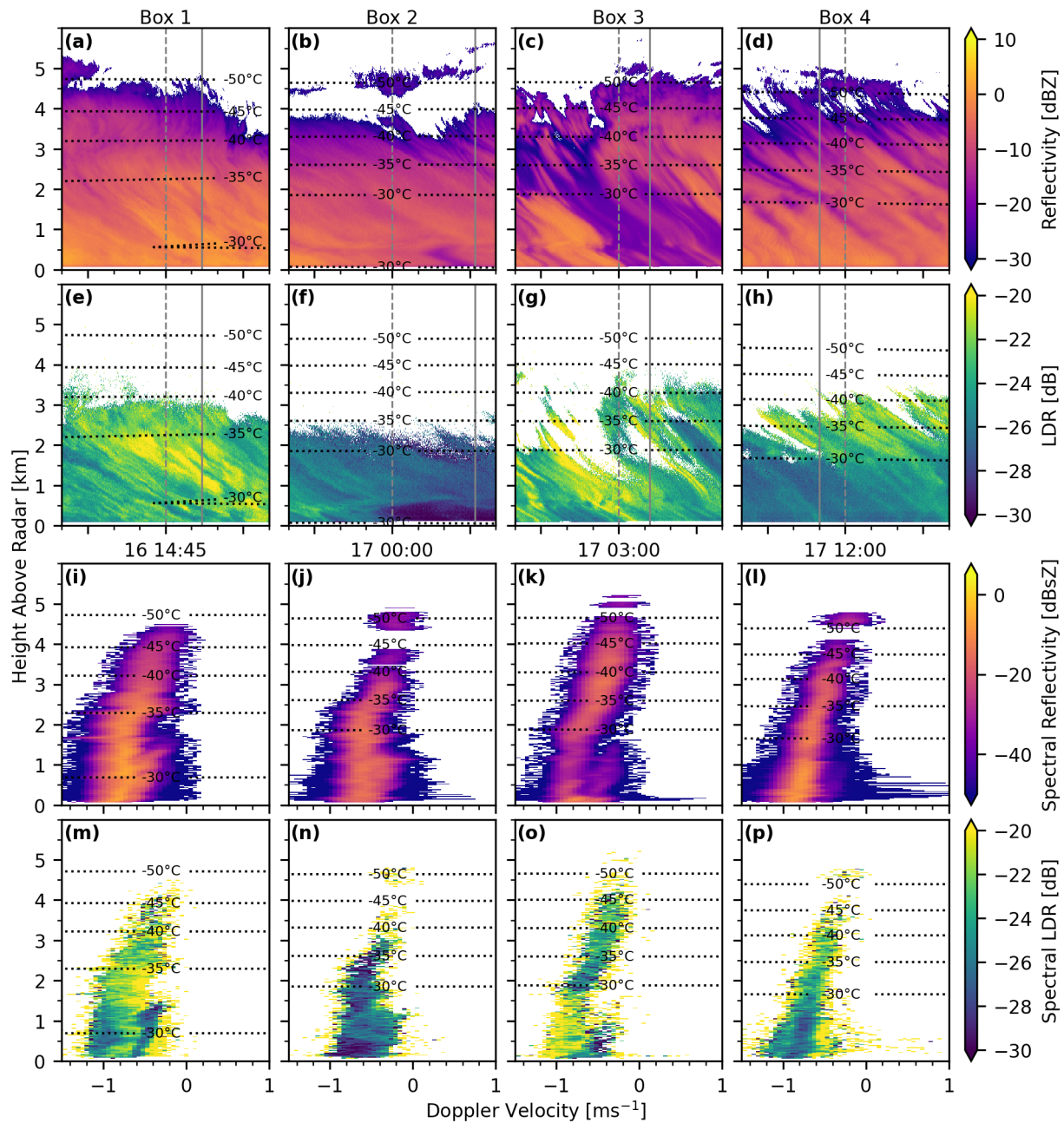


Figure 16. Excerpts of MIRA data at DMC. Each column corresponds to a time of interest discussed in the text. The reflectivity and Linear Depolarisation Ratio LDR time-height sections in (a)-(h) cover 20 minutes either side of the time indicated on either the 16th or 17th of February 2025. Vertical dashed grey lines indicate the time of the spectrograms in (i)-(p), whilst the solid vertical lines indicate the time of the ICE-CAMERA images in Fig. 17 and Appendix C. The temperature contours are from ERA5.



ICE-CAMERA - 16/02/25 14:52 UTC N grains=568



Figure 17. Ice-crystal images from the ICE-CAMERA at Concordia station. This panel shows the ice crystals collected on the plate in the ten minutes preceding the 14:52 UTC on the 16th of February 2025, corresponding to box 1 in Fig. 16.

ice crystals for which the habit cannot be determined. The lack of plate-like crystals suggests that the habit is determined near the cloud top and that growth within the plate regime is limited. This could be due to the lower supersaturation with respect to



ice in the lower part of the cloud (according to the radiosonde in Fig. 15), resulting in a lower rate of vapour deposition in the plate regime. In addition, Bailey and Hallett (2009) report that bullets and rosettes falling into the plate-like regime often only show small changes, such as a thickening of the column elements.

Given the dominance of the column regime in the observed particles, we hypothesise that the variation in LDR arises from changes *within* the column regime. Namely, the proportion of single-column crystals relative to polycrystalline bullet rosettes and small ice crystals. Individual bullets and hollow columns are expected to have a high LDR due to their high aspect ratio, whereas bullet rosettes should have a lower LDR due to their higher sphericity. This has been quantified using scattering simulations at W-band by Sato and Okamoto (2006). Although the results cannot be directly applied to the K_a-band MIRA, we still expect a lower LDR for rosettes than for individual columns.

Turning to the multi-modality identified with peakTree, all the ICE-CAMERA image-collections show a rich spectrum of bullet rosettes, individual bullets, columns and smaller fragments close to the detection limit of the instrument (around 60 µm, Del Guasta, 2022)). It is therefore not unexpected that particle populations with distinct fall-speeds are present, even in periods where rosettes or individual columns are dominant. For time periods in which rosettes are dominant, the secondary peak could correspond to individual bullets and columns. This would lead to a secondary peak with a higher spectral LDR than the main peak. However, in regions where the main peak is dominated by individual bullets and the secondary peak is made up of smaller particles, we might expect the secondary peak to have lower LDR than the primary peak. In fact, both configurations are observed.

Figure 16 shows four 40-minute excerpts ('boxes') of MIRA reflectivity and LDR data, corresponding to the four ICE-CAMERA images in Figures 17, C1, C2 and C3 (Appendix C) respectively. The reflectivity and LDR spectrograms are shown for the central time of each box. The four boxes were chosen to illustrate different representative scenarios of reflectivity and LDR co-variation and spectrogram structure and to attempt to link these with the particle populations observed by the ICE-CAMERA.

Box 1 shows a period of relatively high reflectivity and LDR. The spectrogram shows two peaks. In the faster peak, the fastest particles have the lowest LDR. The slow secondary peak also has a low LDR. The ICE-CAMERA images for this time period include individual columns, bullet rosettes and small crystals. We hypothesise that the majority of the faster peak in the spectrum corresponds to columns, the fastest of which are poly-crystalline rosettes with lower LDR. The secondary peak would correspond to small, spherical ice fragments. The large number of particles and presence of large rosettes and columns leads to a high overall reflectivity. In box 2, the reflectivity and LDR are both low. The ICE-CAMERA images for this time-period show both small ice crystals and polycrystalline or irregular crystals but no individual columns. This results in a particle population in which most particles have a low aspect ratio - and hence low LDR, albeit a large variation in particle size and hence Doppler velocity. Box 3 shows a time period of low reflectivity but high LDR. The ICE-CAMERA images show both columns and small particles but a lack of polycrystalline crystals or rosettes. The spectrogram in this case shows two peaks, the faster of which has the highest LDR. The faster peak likely corresponds to the high aspect ratio columns, which dominate the radar signal and lead to a high overall LDR. Finally, box 4 shows high reflectivity but low or medium LDR. The ICE-CAMERA images show small crystals and short columns but very few polycrystalline rosettes. The spectrogram shows one main peak



455 with relatively low LDR and a less-distinct secondary peak. In the faster peak, the fastest particles have a higher LDR than the slower particles. The hypothesis in this case is that the main peak corresponds to short columns, which despite their columnar habit still have a relatively low aspect ratio and hence low LDR. The largest and fastest of these columns are longer and hence have a higher LDR.

The LDR, reflectivity and spectral structure can therefore be linked to the proportion of columns, polycrystalline rosettes and small ice crystals, modulated by the particle sizes and number concentration. The hypotheses proposed above are qualitative and a more robust attribution would rely on a quantitative investigation of the crystal habits and radar data, which lies beyond the scope of the present study.

Similar to at D85, we question what controls the changing proportions of different crystal habits within the particle population of the moist intrusion. High ice-supersaturations are known to favour the growth of polycrystalline rosettes (Bailey and Hallett, 2009). Therefore, temporal variations in humidity at the cloud top could lead to variations in the proportion of rosettes. Unfortunately, this hypothesis is tricky to validate quantitatively due to the lack of continuous high resolution humidity data throughout the atmospheric column. For example, we identified that ERA5 underestimates the relative humidity with respect to ice compared to the radiosonde for this event (Fig. 15), so cannot be relied upon for identifying other periods of enhanced supersaturation.

470 We are also interested in the factors leading to multi-modality in the particle population. In the peakTree plot (Fig. 14) and spectrograms (Fig. 16), we note that the secondary peak typically appears in the 1 km closest to the ground. In the radio sounding (Fig. 15), this region is quasi-isothermal and shows several low-level jets in the wind speed. This suggests a certain level of mixing, which could lead to secondary particle populations arising through the collisional break-up of bullet rosettes into individual bullets. However, no strong turbulence is observed in the radar MDV or spectral width. In the context of multi-modality, it is important to consider the presence of supercooled liquid water. At DMC, we have more observations to constrain this than at other sites. Both the micro-lidar in the AWACA container and the depolarization lidar of Concordia station (Di Liberto et al., 2024, 2026) show the presence of a small supercooled layer around 1000 m a.g.l. from 22:00 to 23:00 on the 17th, whilst the remainder of the case study does not show strong supercooled layers (not shown). Although the HAMSTRAD radiometer (Ricaud et al., 2010) does show a non-zero liquid water path (not shown), ERA5 suggests that supercooled water is not present over the high plateau (Fig. 6). The lack of visibly-rimed particles in the ICE-CAMERA images also suggests that the quantity of supercooled water is low. We therefore suggest that supercooled water is not a dominant factor controlling the microphysics during this event.

To summarise the observations at DMC, the precipitation is made up of small ice crystals with a columnar habit, corresponding to the temperature at the cloud-top. Varying proportions of polycrystalline bullet-rosettes, individual columns and small ice crystals observed by the ICE-CAMERA are hypothesised to lead to temporal changes in the LDR. As at D85, we suggest that changing thermodynamic conditions can influence the dominant particle habit in a distribution of particles.

It is interesting to note that in the extremely cold environment at DMC, significant variation in particle habit and size is still observed. A tentative analogy can be drawn to cirrus clouds, which occupy a similar temperature regime ($< 235\text{ K}/-38^{\circ}\text{C}$, Krämer et al., 2016). Such an analogy has already been made by Vignon et al. (2022), albeit for ice fogs formed in-situ



at Concordia, rather than coastal-driven cloud systems. The review of Lawson et al. (2019) highlights the diversity of particle shapes in cirrus clouds and the dominance of bullet rosettes and polycrystals, as we observe at DMC during the moist intrusion. Doppler spectrum observations of mid-latitude cirrus clouds are challenging to make, since cirrus lie at altitude ranges beyond the sensitivity limit of ground-based cloud radars. The ongoing AWACA measurements have the potential to provide insights into cloud micro-physics in cirrus conditions.

9 Discussion and Conclusion

This work presents a case study of a moist intrusion from a coastal cyclone in the vicinity of DDU moving onto the high plateau in East Antarctica. The passage of the intrusion – as revealed by Lagrangian trajectory analysis – over four observational stations allows the micro-physical processes of the resulting snowfall to be investigated. Analysis of the multi-frequency, polarimetric and spectral radar data along the transect reveals that, despite the precipitation being driven by the same weather system, there are strong differences in the micro-physical processes controlling the growth of ice particles between the coast and high plateau.

At the two sites closest to the coast (D17 and D47), the presence of supercooled liquid water, significant upwards air motion and turbulence leads to the growth of large particles through riming and aggregation. The temperature conditions at the coastal sites are not far-removed from mid-latitude winter temperatures and as a result these processes and their signatures are familiar to the radar community (eg. Grazioli et al., 2015; Kalesse et al., 2016; Teisseire et al., 2025). In Antarctica, riming has been proposed to explain the observed higher fall-speed by MRRs at DDU compared to at PE (Durán-Alarcón et al., 2019). Aggregation and riming were inferred from DFR and MDV respectively by Alexander et al. (2023) for snowfall associated with coastal cyclones at Davis station. At DDU, riming and aggregation were observed by an X-band radar and snowfall camera in the APRES3 campaign (Grazioli et al., 2017a). The link between ascending air streams in cyclone systems and enhanced riming through the availability of supercooled water and turbulence is highlighted in the case study of Gehring et al. (2020). Although the ascent observed in our case study is lower than in mid-latitude WCBs, we still expect that it enhances the availability of supercooled droplets and hence the probability of riming. To accurately simulate snowfall in Antarctica, models should be able to resolve dynamical structures above the coastal slope and account for their impact.

As the intrusion progresses inland over D85 and DMC (477 and 1103 km from the coast respectively), the particle size decreases, as revealed by the radar reflectivity and DFR. A decrease in particle size between the Antarctic coast and interior has previously been indirectly documented by studies of snowfall-rate quantification from radar reflectivity. Namely, the multiplier in the power law between snowfall rate and reflectivity is known to depend on the typical snowflake size (eg. Schoger et al., 2021) and the optimal multiplier is smaller at the inland site PE than at DDU (Souverijns et al., 2017). This difference was attributed to the drier conditions at PE.

Two factors are expected to influence the decreasing particle size observed in the moist intrusion in this study. Firstly, the depleted water content as the intrusion moves inland, arising from the high snowfall rates close to the coast. Secondly, decreased efficiency of ice crystal growth processes (such as riming, dendritic growth and aggregation) at very low temperatures. The



radar signatures at D85 and DMC reveal changes in the dominant habits of the small ice particles, which we suggest are principally controlled by the temperature and ice supersaturation at the top of the cloud. The conclusions at the inland sites are more tentative than at the coastal sites, due to the lack of Doppler cloud radar observations in similar cold-temperature conditions. In particular, the attribution of peaks in the observed multi-modal Doppler spectra remains speculative.

Here we emphasise the difference between the dominant processes inferred in the case study at the coastal and inland sites. However, it is important to note that each site shows similarities to each adjacent site, such that there is no strong boundary between ‘coastal’ and ‘inland’ conditions, but rather a smooth transition in the dominance of different processes as the moist intrusion travels inland. For example, although riming is assumed to be present throughout the event at D17, it is still evident during shorter time periods at D47 (Section 6). Indeed, the investigation of this gradient in process occurrence motivates the installation of the remote temporary stations between the manned stations at the extreme ends of the transect.

The deployment of a transect of temporary stations collecting multi-frequency, polarimetric, spectral radar data has permitted the investigation of snowfall resulting from an Antarctic moist intrusion in February 2025. The micro-physical processes leading to the growth of precipitating snowflakes from the water vapour within the intrusion differ along the transect according to the thermodynamic conditions and dynamical effects such as coastal uplift. The planned three-year time series of observations along the AWACA transect will allow further insights into micro-physical processes over multiple events and in different seasons.

Code and data availability. The radar data will be made available in the context of the AWACA project and is available before then from the contact authors by request. ERA5 data can be downloaded from the Copernicus climate data store (<https://doi.org/10.24381/cds.bd0915c6>, last access 12/06/2026). Radiosonde data from Dumont d’Urville station can be downloaded from the Météo-France website (https://donneespubliques.meteofrance.fr/?fond=produit&id_produit=97&id_rubrique=33, last access 12/06/2026). Radiosonde data for Concordia station (Grigioni et al., 2022) can be requested from https://www.climantartide.it/dataaccess/RDS_CONCORDIA/index.php?lang=en (last access 12/06/2026).

The code for the peakTree algorithm (Radenz et al., 2019) can be downloaded from <https://github.com/martin-rdz/peakTree> (last access 12/06/2026). The code for the LAGRANTO algorithm (Sprenger and Wernli, 2015) can be accessed by request to Michael Sprenger.

This work makes use of the following collaborative scientific software projects for data analysis and display: xarray (Hoyer and Hamman, 2017), MetPy (May et al., 2022), Cartopy (MetOffice, 2010), Matplotlib (Hunter, 2007).

Appendix A: Calibration

This section describes the relative and absolute calibration of the radars on the AWACA transect. The inter-calibration of the three radars at each site (MRR, MIRA and BASTA, see section 2.1) is important for the calculation of the dual frequency ratios DFR_{K-K_a} and DFR_{K_a-W} . We are interested in the component of the DFR resulting from differences in the scattering regime (DFR_{scat}) at two frequencies: as the particle size approaches the wavelength of the incident radiation, the rate of increase of reflectivity with particle size decreases as the scattering exits the Rayleigh regime. Since the decrease occurs first at the



Table A1. Maximum two-way path integrated attenuation at 5 km range (5 km a.g.l.) due to atmospheric gases during the case study, in dB. Calculated using PAMTRA (Mech et al., 2020).

	MRR	MIRA	BASTA
D17	0.54	0.36	1.09
D47	0.35	0.25	0.65
D85	0.19	0.17	0.36
DMC	0.12	0.14	0.24

smaller wavelength, the difference in reflectivities can give an indication of the particle size. However, there are additional contributions to the measured DFR at any range gate. Similar to the approach of von Terzi et al. (2022), we can write

$$\text{DFR} = \text{DFR}_{\text{scat}} + \text{DFR}_{\text{gas}} + \text{DFR}_{\text{hard}} + \text{DFR}_{\text{hydro}}, \quad (\text{A1})$$

where DFR_{gas} represents the differential attenuation due to atmospheric gases, DFR_{hard} represents hardware differences in the radar, including a miscalibration of the reflectivity, and $\text{DFR}_{\text{hydro}}$ is differential attenuation due to hydrometeors, particularly liquid particles. The estimation and correction of these contributions is described in the order in which they are performed in sections A1, A2 and A3 respectively. Finally, consideration is given to the inter-calibration of the radars at different sites (A4).

A1 Gas Attenuation Correction

Atmospheric gases lead to attenuation of the radar signal. This attenuation increases with range and is larger for higher frequencies. The Passive and Active Microwave radiative TRAnsfer tool (PAMTRA, Mech et al., 2020) was used to calculate the two-way path-integrated attenuation arising from oxygen, nitrogen and water vapour, using the gas absorption model of Rosenkranz (1993, 1998). This corresponds to DFR_{gas} in Eq. (A1). The required input profiles of temperature, relative humidity and pressure were taken from ERA5. Table A1 gives the maximum two-way path integrated attenuation due to atmospheric gases at 5 km range for the radars at the four sites. The gas attenuation is greatest for the W-band BASTA and for the sites near the coast, where the concentration of water vapour is higher. For each site, the gas attenuation is also greater after the arrival of the precipitating air mass (on the 16th for D17, D47 and D85, and the 17th for DMC).

A2 Inter-calibration between radars in the same OPU

The largest contribution to the measured DFR arises from the constant offset in the reflectivities measured by the different radars (DFR_{hard}). This was calculated after gas attenuation correction using the calibration transfer methodology of Jorquera et al. (2023). The calibration coefficient (reflectivity offset) between two radars is calculated using a linear regression on filtered reflectivity data, focusing on the reflectivity range for which the two reflectivities most closely follow a 1:1 slope. An example of the MIRA-BASTA inter-calibration is shown in Fig. A1. For each site and for each pair of radars, multiple calibration periods

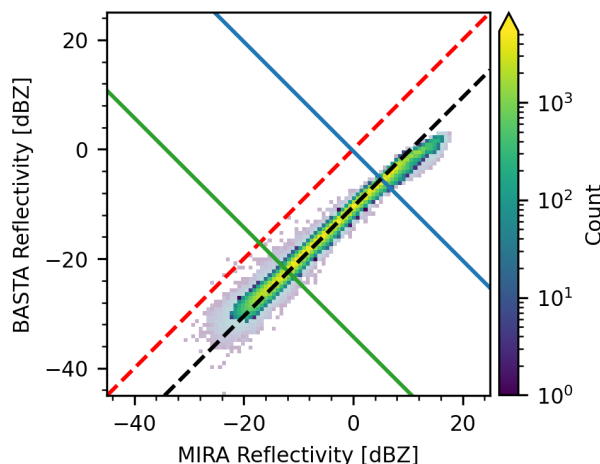


Figure A1. An example of the radar inter-calibration method of Jorquera et al. (2023), applied between the MIRA and BASTA radars at D47 between 1000 and 9000 m a.g.l. for 00:00 UTC to 03:00 UTC on the 16th. The red dashed line shows the 1:1 line. Data masked by the density mask is shown as partially transparent. The green and blue solid lines bound the region selected for the best linear fit. The final linear fit is shown with a black dotted line and gives a calibration coefficient of 10.5 ± 0.7 dBZ for this time period.

of several hours were used to estimate the calibration constant CC and its uncertainty σ_{CC} . The estimates from individual time periods were combined to give the final estimates in Table A2. The MRR-BASTA offset is more uncertain than the other radar pairs since the regions of typically large, precipitating particles detected by the MRR are likely to be those for which the BASTA is already outside the Rayleigh regime. Hence it is difficult to find time periods for which the linear relation between the two reflectivities meets the statistical criteria set out by Jorquera et al. (2023). However, the closure of the three estimates at D17, D47 and D85 (the residual calibration constant is zero within the uncertainty) suggests that the technique is robust. At DMC, the low sensitivity of the MRR meant that no suitable period for inter-calibration were found. However, the MRR data was not used at DMC during the case study. It can be seen that the offset of the BASTA can be very large, whereas the MRR and MIRA show only a small offset. This is likely due to the absolute calibration of the BASTA calculated during testing in Europe not being valid after the installation in Antarctica (Julien Delanoë, personal correspondence). In practice, the MIRA is used as a reference and the BASTA is corrected. The very small MIRA-MRR offset is not applied as it likely lies within the uncertainty of the measured reflectivities. Note that the inter-calibration estimate is specific to the time period of the case study. Changes in the transmit power and receiver sensitivity can lead to changes in the inter-calibration over time.

590 A3 Liquid attenuation correction

Estimating DFR_{hydro} is difficult due to the lack of information on the vertical distribution of liquid hydrometeors. This information is not available from the instruments in the AWACA OPU. However, after gas attenuation correction and inter-calibration, we noticed that, at D17, a significant DFR_{K_a-W} signal remained near the cloud top after the start of precipitation. The DFR



Table A2. Calibration constants (reflectivity offsets) in dB between the three radar pairs at each site, as estimated using the methodology of Jorquera et al. (2023). The residual gives the sum of the first two offsets minus the third offset. For DMC, the low sensitivity of the MRR meant that no calibration periods close to the case study met the statistical criteria for inter-calibration set out in Jorquera et al. (2023).

	MIRA-BASTA	MRR-MIRA	MRR-BASTA	Residual
D17	4.5 ± 0.5	0.0 ± 0.4	4.3 ± 0.9	0.3 ± 1.1
D47	10.4 ± 0.3	0.6 ± 0.8	11.6 ± 0.9	-0.7 ± 1.2
D85	18.6 ± 0.3	1.1 ± 0.4	19.4 ± 0.4	0.4 ± 0.6
DMC	27.1 ± 0.4			

signal near the cloud top is expected to be close to 0 dB, since small ice particles can be assumed to be Rayleigh scatterers even in the W-band. We therefore attributed this offset to attenuation by liquid hydrometeors at W-band. This attribution is supported by the fact that the offset only appears after the precipitating cloud reaches the surface late on the 15th, by the fact that it only appears close to the coast and that it corresponds to periods of elevated CLWC in ERA5 (up to a maximum of 0.17 g m^{-3} on the 16th). A similar attenuation signal could arise from accumulation of wet snow on the radomes of the OPU. However, webcam images of the OPU roof (not shown) show that very little snow settled on the radomes.

To account for this offset, we used the method of Dias Neto et al. (2019) (see also von Terzi et al. (2022)) to calculate the mean difference in reflectivities at the top of the cloud. Using the MIRA as a reference, the mean difference in reflectivities between the BASTA and MIRA was calculated using a 15-minute rolling window. Only regions near the cloud top (range greater than 3.5 km) and with low reflectivities in the K_a band ($-30 \text{ dBZ} < Z_{eK_a} < 5 \text{ dBZ}$) were used. The reflectivity range is slightly extended compared to Dias Neto et al. (2019) due to the low sensitivity of the BASTA. As in (Dias Neto et al., 2019) and (von Terzi et al., 2022), regions with fewer than 300 valid reflectivity pairs, a correlation of less than 0.7 between the two reflectivities or a DFR variance of more than 2 dB^2 are excluded. These regions are filled with a linear interpolation between valid time points. The calculated offset $\text{DFR}_{\text{hydro}}$ is applied uniformly over the whole profile. This leads to an underestimation of DFR_{scat} at low ranges, since we overcompensate for liquid attenuation at low ranges.

Note that we apply the method of Dias Neto et al. (2019) after applying the offset calculated in Appendix A2 (DFR_{hard}). The additional offset is therefore expected to be zero when liquid attenuation is not significant. An alternative approach would be to calculate the combined $\text{DFR}_{\text{hard}} + \text{DFR}_{\text{hydro}}$ with the approach of Dias Neto et al. (2019). However, since the offset resulting from the pre-deployment calibration of the BASTA was known to be large, we preferred to estimate this independently.

Figure A2 shows the calculated offset for D17 and the DFR_{K_a-W} before and after the offset is applied. It can be seen that the offset effectively reduces the DFR near the cloud top but does not change the overall structure of the DFR signal in time-height space. Although the $\text{DFR}_{\text{hydro}}$ was also estimated for the other sites, it was found to oscillate close to 0 dB and was therefore not applied. This supports the estimates of DFR_{hard} calculated in Appendix A2 and suggests that liquid attenuation is negligible at D47, D85 and DMC. This is consistent with ERA5, which suggests the CLWC stays well below 0.1 g m^{-3} at

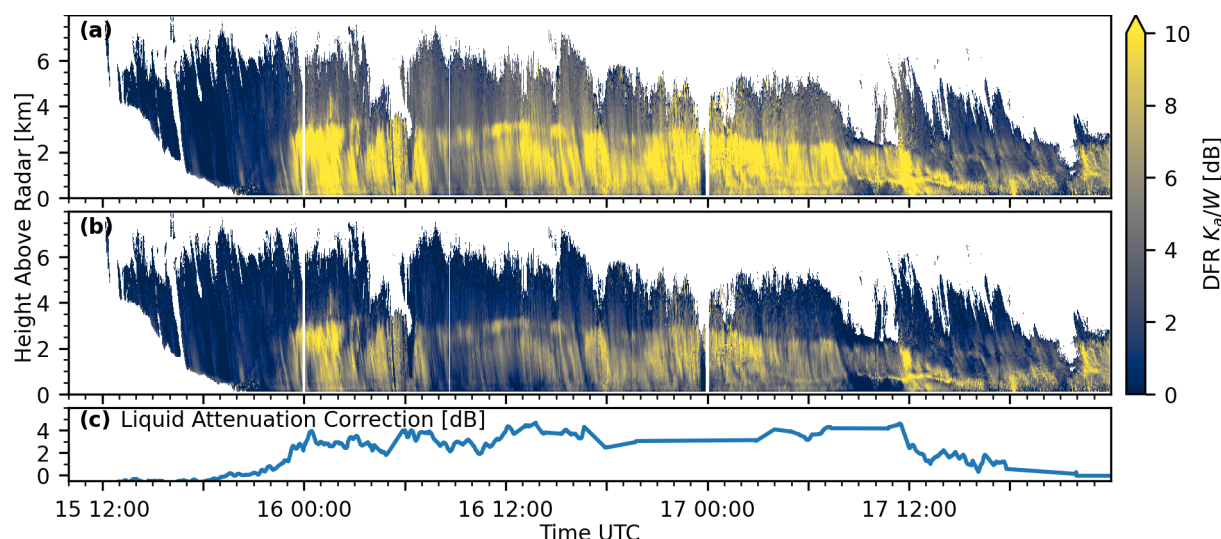


Figure A2. The Dual Frequency Ratio (DFR) between the MIRA and BASTA at D17, before (a) and after (b) liquid attenuation correction to the BASTA reflectivity. The magnitude of the correction is shown in (c).

D47, D85 and DMC for the duration of the event. Due to the uncertainty inherent in the liquid attenuation correction, the DFR is primarily used qualitatively in this study.

620 A4 Inter-calibration between different OPUs

The inter-calibration of the radars in different OPUs is important to check that reflectivity differences along the transect can be attributed to differences in the physical properties of the precipitation rather than instrument biases. However, this inter-calibration is challenging as the radars at different sites do not measure the same profile of clouds and precipitation.

A first approach to estimate the inter-calibration uses the reflectivity-range histograms (Fig. A3). This is a similar approach
625 to that used by Ferrone et al. (2023) for comparing multiple MRR radars. The position of the minimum reflectivity measured at each range (sensitivity curve, on the left hand side of the histogram) gives an approximation for the reflectivity offset between radars of the same frequency at different sites. It can be seen that whilst the MIRA and MRR radars have comparable sensitivity curves at the four sites, the histograms for the BASTA are offset. The offset between sites agrees well with the MIRA-BASTA offset at each site (Table A2), suggesting that the MIRA can be taken as a reference whereas the BASTA should be corrected
630 at each site.

This hypothesis is supported by a statistical comparison of the BASTA reflectivities to those of the 94 GHz cloud profiling radar of the EarthCARE satellite (Illingworth et al., 2015) performed by the LATMOS laboratory (Feuillard et al., 2026). An offset of 10 dB at D47 (Nathan Feuillard and Felipe Toledo Bittner, personal correspondence) agrees with the 10 dB offset calculated relative to the MIRA (Table A2).

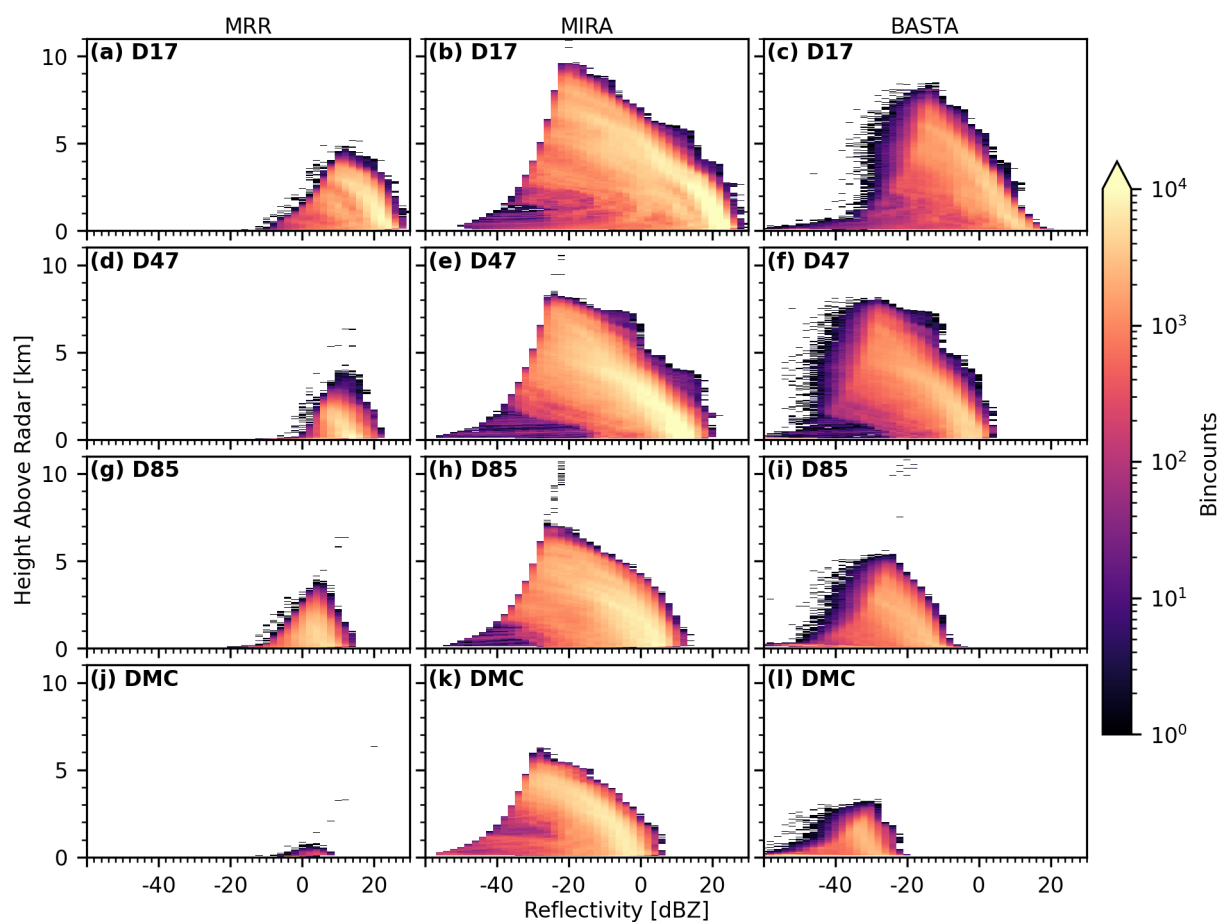


Figure A3. Histograms of reflectivity for the three radars at each site of the AWACA transect. The reflectivity plotted is the raw reflectivity before attenuation or inter-calibration corrections.



635 Appendix B: PAMTRA simulations for dual frequency signatures

Figure B1 shows simulations of DFR_{K_a-W} and DFR_{K-K_a} performed using PAMTRA (Mech et al., 2020) and the Self-Similar Rayleigh-Gans (SSRG) framework. The simulation is run for a single particle of a given size. The size of the particle is then increased. The results show that for particles starting with a size of 4-5 mm, it is possible to have an increase in DFR_{K-K_a} before an increase in DFR_{K_a-W} with further particle growth. This effect is discussed in Section 4.

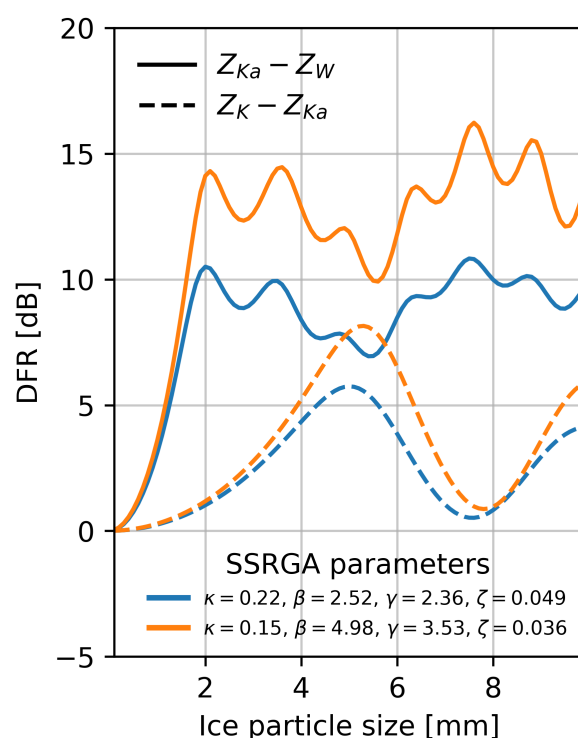


Figure B1. PAMTRA simulations of the dual frequency ratio (DFR) for two sets of representative Self-Similar Rayleigh-Gans (SSRG) parameters. The SSRG parameters are taken from table 2 in (Billault-Roux and Berne, 2025) and are based on the snowScatt model (Ori et al., 2021).

640 Appendix C: ICE-CAMERA Images

Figures C1, C1 and C3 show the ice crystal images from the ICE-CAMERA corresponding to boxes 2, 3 and 4 in Fig. 16 (Section 8) respectively.



ICE-CAMERA - 17/02/25 00:16 UTC N grains=223

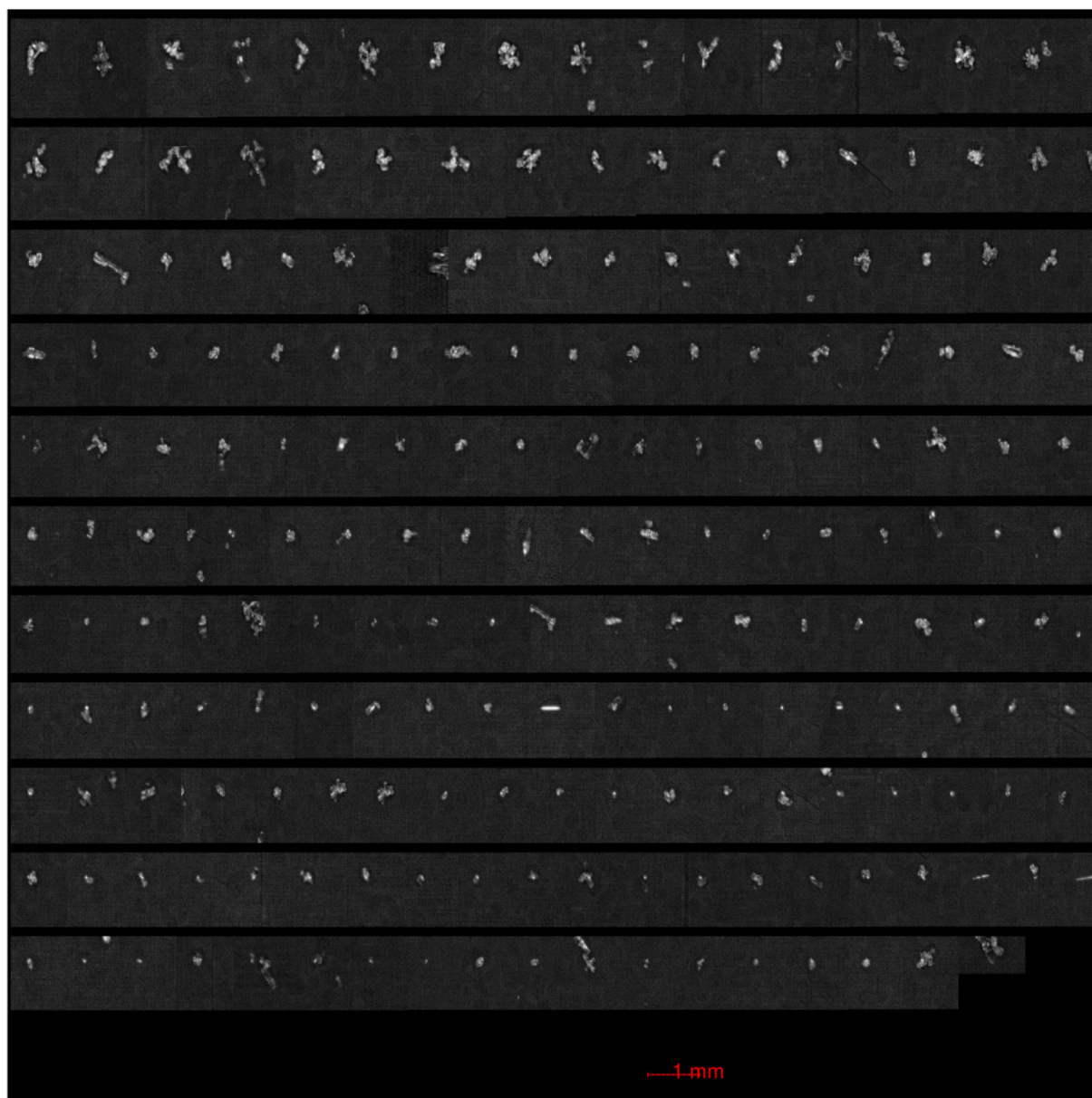


Figure C1. Ice-crystal images from the ICE-CAMERA at Concordia station. This panel shows the ice crystals collected on the plate in the ten minutes preceding the 00:16 UTC on the 17th of February 2025, corresponding to box 2 in Fig. 16.



ICE-CAMERA - 17/02/25 03:06 UTC N grains=780

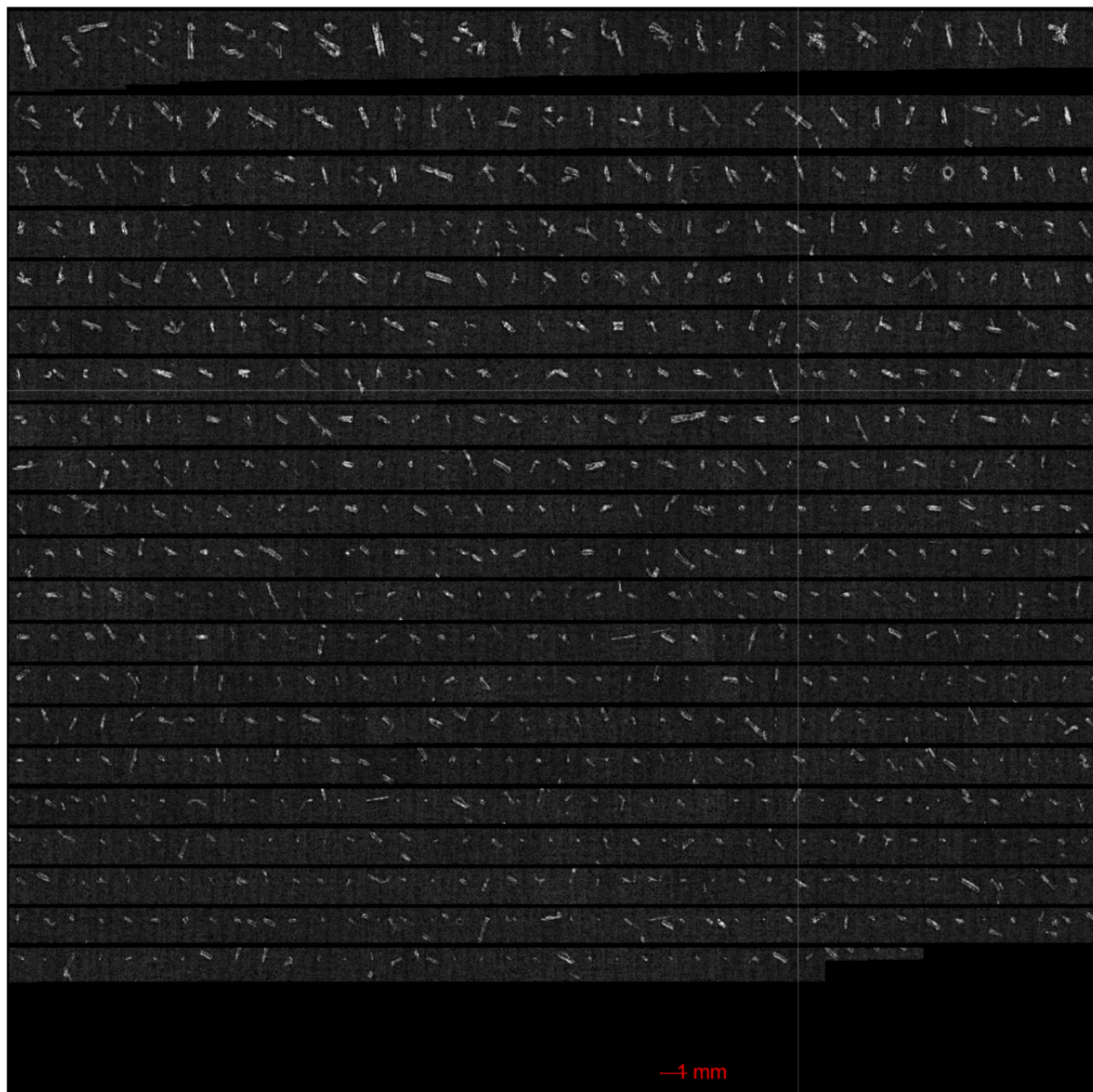


Figure C2. Ice-crystal images from the ICE-CAMERA at Concordia station. This panel shows the ice crystals collected on the plate in the ten minutes preceding the 03:06 UTC on the 17th of February 2025, corresponding to box 3 in Fig. 16.



ICE-CAMERA - 17/02/25 11:55 UTC N grains=239

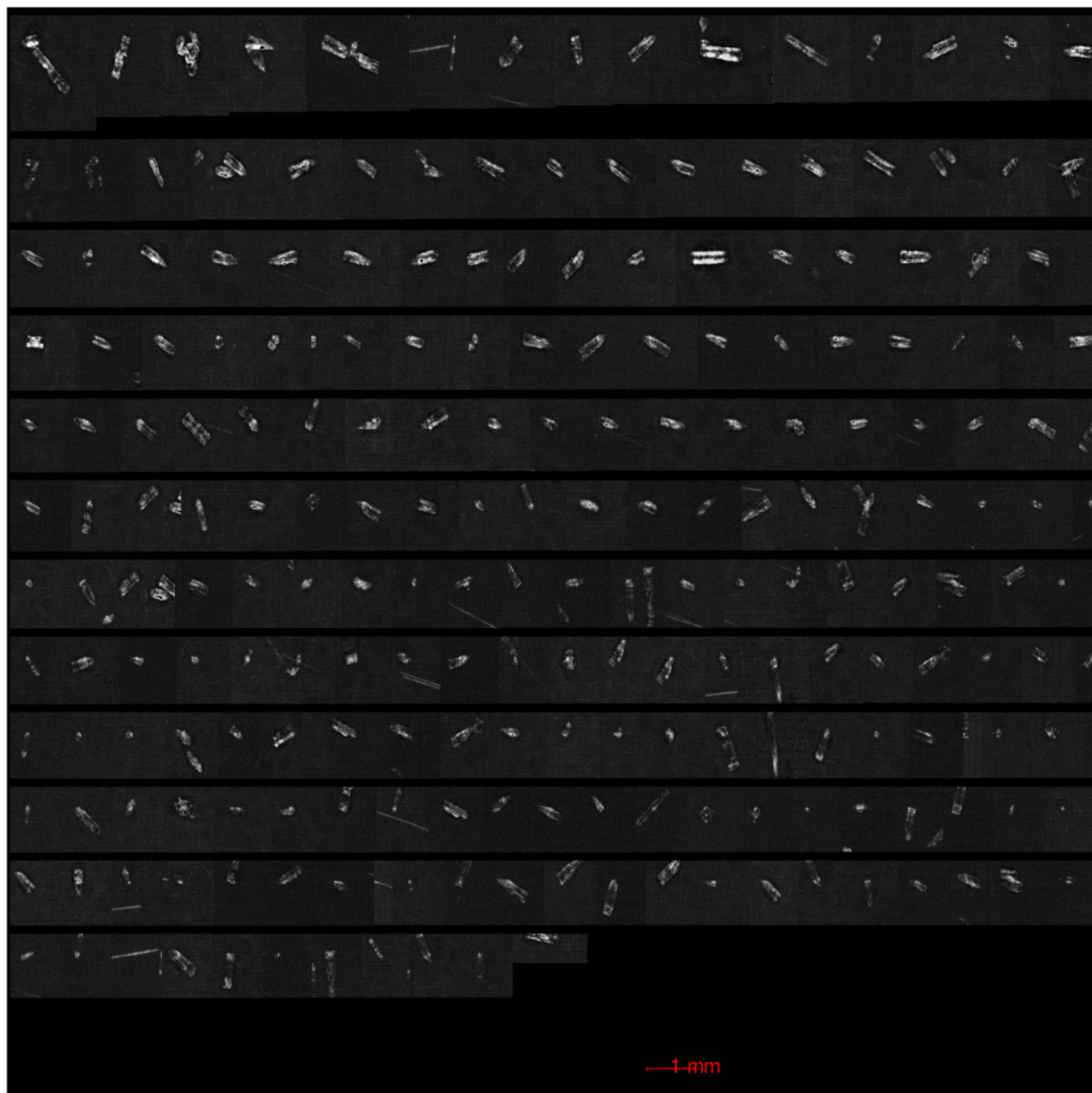


Figure C3. Ice-crystal images from the ICE-CAMERA at Concordia station. This panel shows the ice crystals collected on the plate in the ten minutes preceding the 11:55 UTC on the 17th of February 2025, corresponding to box 4 in Fig. 16.



Author contributions. HC and AB conceived the study. HC processed and analysed the radar measurements and auxiliary data under the supervision of AB. JD managed the BASTA radar installation and operation and processed the BASTA data. MDG installed the ICE-
645 CAMERA at Concordia and processed the ice crystal images. HC wrote the paper with contributions from AB, JD and MDG.

Competing interests. The authors declare that they have no conflict of interest.

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