

1 **Fingerprints of a riming event on cloud radar Doppler**
2 **spectra: Observations and Modeling**

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2 **Abstract**

3 Radar Doppler spectra measurements are exploited to study a riming event when precipitating
4 ice from a seeder cloud sediments through a supercooled liquid water (SLW) layer. The focus
5 is on the “golden sample” case study for this type of analysis based on observations collected
6 during the deployment of the Atmospheric Radiation Measurement Program’s (ARM) mobile
7 facility AMF2 at Hyytiälä, Finland during the BAECC (Biogenic Aerosols – Effects on 
8 Clouds and Climate ~~Snowfall Experiment~~) field campaign. The presented analysis of the
9 height evolution of the radar Doppler spectra is a state-of-the-art retrieval with profiling cloud
10 radars in SLW layers beyond the traditional use of spectral moments. Dynamical effects are
11 considered by following the particle population evolution along slanted tracks that are caused
12 by horizontal advection of the cloud under wind shear conditions. In the SLW layer, the
13 identified liquid peak is used as an air motion tracer to correct the Doppler spectra for vertical
14 air motion and the ice peak is used to study the radar profiles of rimed particles. A 1D steady-
15 state bin microphysical model is constrained using the SLW and air motion profiles and cloud
16 top radar observations. The observed radar moment profiles of the rimed snow can be
17 simulated reasonably well by the model, but not without making several assumptions about
18 the ice particle concentration and the relative role of deposition and aggregation. This
19 suggests that in-situ observations of key ice properties are needed to complement the profiling
20 radar observations before process-oriented studies can effectively evaluate ice microphysical
21 parameterizations.

22

23 **1 Introduction**

24 Mixed-phase clouds are ubiquitous, long-lived and cover extended areas (e.g., Shupe et al.,
25 2008; Zhang et al, 2010; Kanitz et al., 2011). However, the factors governing the formation,
26 maintenance and dissipation of mixed-phase clouds are poorly understood and consequently
27 not well represented in weather and climate models (Cantrell and Heymsfield, 2005; Lebo et
28 al., 2008; Barrett et al., 2010). The complex interaction between atmospheric vertical motions,
29 aerosol particles, water vapor, liquid water, and ice determine the radiative and microphysical
30 properties of mixed-phase clouds to a large extent (Gregory and Morris, 1996). Microphysical
31 processes such as water vapor diffusion, collision, coalescence, aggregation, and riming are

1 controlled by the variable mass ratio between liquid water and ice (Pruppacher and Klett,
2 1997). The ongoing increase in the temporal and spatial resolution of numerical models
3 suggests that cloud microphysical processes will be modeled in ever more detail in the
4 coming decades (Klein et al., 2013). In that context, the development of process-level
5 understanding has been found to be a key for success in addressing the complicated nature of
6 mixed-phase clouds and improving their representation in numerical models (Morrison et al.,
7 2012).

8 Mixed-phase clouds pose a serious observational challenge due to the difficulty of identifying
9 the presence of SLW layers embedded in cloud regions dominated by ice (Luke et al., 2010).
10 Existing mixed-phase cloud classifications are highly uncertain and lead to a
11 misrepresentation of these clouds in models (Illingworth et al., 2007). Moving beyond the
12 detection of SLW layers and into process-oriented studies (e.g., riming) requires synergetic
13 observations with cloud Doppler radars and microwave radiometers (MWR) in combination
14 with backscatter and Doppler lidars (e.g., Verlinde et al., 2013). As highlighted in Kollias et
15 al. (2007a) spectral Doppler information is expected to be one of the main tools for future
16 observational studies on cloud microphysics (see Section 2.3 for details).

17 Here, a 35-GHz cloud Doppler radar is used in synergy with a microwave radiometer to
18 identify and characterize a SLW layer within a mixed-phase cloud and its effect on the cloud
19 microphysics. The recorded radar Doppler spectra are bimodal, thus comprised of a liquid and
20 an ice spectral peak (e.g., Shupe et al., 2004). As in Shupe et al. (2004), the vertical air motion
21 within the SLW layer is retrieved from cloud radar Doppler spectra. The liquid peak radar
22 reflectivity is used to retrieve the SLW profile while the spectral peak associated with typical
23 ice and snow terminal velocities is used to detect and follow the evolution of riming. The
24 temporal (height) evolution of the radar Doppler spectrum is analyzed along slanted fall
25 streaks from cloud top to cloud base to optimally follow the particles' history in order to gain
26 insight into microphysical processes occurring in different layers of the mixed-phase cloud
27 (Marshall, 1953; Hogan and Kew, 2005).

28 This study illustrates the objective steps in identifying the impact of a microphysical process
29 (riming) on radar observations (fingerprints) and the steps required to analyze a multi-sensor
30 dataset containing radar Doppler spectra. While the implementation of the aforementioned
31 retrieval and analysis technique is valuable, this is not purely a retrieval effort. The main
32 question this study aims to address is to what extent such process-oriented studies (i.e.,


1 fingerprinting studies) can be used to evaluate existing riming efficiency parameterizations
2 (e.g., Hall 1980; Cober and List 1993; Lohmann 2004). To accomplish this, a 1D steady-state
3 bin microphysics model is used to model the riming event. Restating that this is not a retrieval
4 contribution, the goal is not to reproduce the evolution of riming ice spectra peak moments
5 but rather to assess whether the observations can sufficiently constrain other parameters and
6 factors that can affect the model output.

7 The structure of the paper is as follows. The data, instrumentation and background of radar
8 Doppler spectra processing are introduced in Section 2. Section 3 gives a detailed analysis of
9 the snowfall case study including a description of the synoptic situation (Section 3.1), the in-
10 situ observations (Section 3.2), the cloud radar- and MWR observations as well as the fall
11 streak tracking technique and the evolution of the cloud radar Doppler spectrum (Section 3.3
12). In Section 3.4 the 1D microphysical bin model used to reproduce the rimed mode radar
13 moments is described and a comparison of observations with model results is discussed. A
14 summary and conclusions are provided in Section 4.

15

16 **2. Data, Instrumentation and Doppler Spectra Processing**

17 **2.1 BAECC Field Campaign Overview**

18 From February 1 to September 12, 2014 the genic Aerosols - Effects on Clouds and
19 Climate (BAECC) field experiment (Petäjä, 2013) - a joint project of the University of
20 Helsinki, the Finnish Meteorological Institute, and the US Department of Energy (DOE)
21 Atmospheric Radiation measurement (ARM) program - took place in the Boreal forest of
22 Southern Finland. For that purpose, extensive remote-sensing and in-situ instrumentation was
23 installed ~~at the~~ Hyytiälä field station of the University of Helsinki located at 61°50'37.114'' N
24 and 24°17'15.709'' E, 150m above sea level (~~Hari and Kulmala, 2005~~). Within that frame, an
25 intensive observation period (IOP) focusing on winter precipitation (BAECC-Snowfall
26 Experiment (SNEX)) organized in collaboration with the National Aeronautics and Space
27 Administration (NASA) Global Precipitation Measurement (GPM) ground validation program
28 and Colorado State University was conducted from February 1 to April 30, 2014.

29

30 **2.2 Instrumentation**

1 The second ARM Mobile Facility (AMF2), consisting of an extensive suite of remote-sensing
2 instruments such as a Ka-band ARM Zenith-pointing Radar (KAZR), a W-, Ka- and X-band
3 Scanning ARM Cloud Radar (Kollias et al., 2014), a Micropulse Lidar (MPL), a High
4 Spectral Resolution Lidar (HSRL) and a two-channel MWR (Cadeddu et al., 2013) was
5 deployed at the observation site. The lidars are used for detection of cloud base height and
6 cloud particle phase, the radars for characterization of cloud- and precipitation microphysics,
7 and the MWR for determination of column-integrated amounts of liquid water and water
8 vapor. For this study, data from the Ka-band ARM Zenith-pointing Radar (KAZR) operating
9 at 35GHz, as well as the MWR are used.

10 Ground-based in-situ sensors included a Particle Imaging Package (PIP), which is a new
11 version of the Snow Video Imager (cf. Newman et al., 2009). Pluvio weighing gauges were
12 employed to measure precipitation rate and snowfall accumulation. Pluvios were also used in
13 combination with the PIP for determination of total particle concentration, particle size
14 distribution (PSD), and particle terminal fall velocities from which fall velocity-size relations
15 were derived at high temporal resolution.

16 For a detailed description of the measurement site set-up as well as the in-situ and remote-
17 sensing instrumentation and data processing please refer to Kneifel et al., 2015. In addition to
18 the mentioned instrumentation, radiosondes were launched four times daily for profiling of
19 the atmospheric state variables.


20

21 **2.3 Doppler spectra processing**

22 The mean Doppler velocity from profiling radars has been used in past studies to detect and
23 study riming. Initially, Weiss and Hobbs (1975) distinguished ice crystal growth by riming
24 from growth by water vapor deposition by observing the different rates of change of mean
25 Doppler velocity with height. Mosimann (1995) used a vertically pointing Doppler radar in
26 combination with snow crystal in-situ measurements to quantitatively determine an index of
27 the degree of riming in stratiform precipitation, a technique which has been used in further
28 studies (e.g., Borys et al., 2003; Baschek et al., 2004).

29 Today, advancements in signal processing and radar technology and decreasing storage costs
30 have enabled the routine recording of the full radar Doppler spectrum. The radar Doppler
31 spectrum often contains unique signatures that can be used to retrieve cloud microphysics and

1 dynamics (Kollias et al., 2007b). In particular, the presence of liquid cloud droplets in the
2 radar sampling volume allows use of the Doppler spectra peak of the liquid particles to derive
3 the mean vertical air motion of the sampling volume. This approach is based on the
4 assumption that the terminal velocity of small cloud droplets is negligible compared to typical
5 vertical air motions in clouds (Kollias et al., 2001). Thus, the location of the peak caused by
6 liquid droplets in the Doppler spectrum can act as a tracer for vertical air motion.

7 This implies that in SLW layers also containing ice particles, if the liquid spectral peak does
8 not significantly overlap the ice peak in velocity, then the cloud dynamics (vertical air motion 
9 and eddy dissipation rate) can be retrieved (Shupe et al., 2004; Kollias et al., 2001). The
10 potential of using multi-modal cloud radar Doppler spectra for characterizing the liquid- and
11 ice-phase components in mixed-phase clouds has been previously demonstrated (e.g., Shupe
12 et al., 2004; Luke et al., 2010; Luke and Kollias, 2013, Rambukkange et al., 2011; Verlinde et
13 al., 2013; Yu et al., 2014).

14 Here, in addition to the objective detection and analysis of the SLW spectral peaks, the
15 temporal evolution of the radar Doppler spectrum is analyzed along slanted fall streaks from
16 cloud top to cloud base to gain insight into microphysical processes occurring in different
17 layers in mixed-phase clouds. As already highlighted in Marshall (1953), this is necessary in
18 situations when vertical wind shear is observed. Under these conditions, following the particle
19 evolution along straight vertical paths is not suitable for detailed fingerprinting studies. To
20 track radar moments in cirrus, this technique has been refined in Hogan and Kew (2005).
21 Similar to their approach, we do not simply follow the fall streaks in the observations but
22 compare them with simulated fall streaks by using the horizontal wind profile and Doppler
23 fall velocity. In this way we can ensure consistency between the expected fall streak shape,
24 which is solely based on dynamics and particles' fall velocity, and the observations. This
25 approach helps to avoid a subjective and potentially false identification of a fall streak in the
26 observations that might be caused, e.g., by different generating processes and levels or
27 directional wind shear that would hamper the derivation of particle history along the fall
28 streak. To our knowledge, this is the first study in which the evolution of the full Doppler
29 spectrum along *slanted* fall streak paths is analyzed.

30 Finally, the SLW radar reflectivity is extracted from the SLW spectral peak and used to derive
31 the profile of SLW content within the SLW layer while the ice rimed spectral peak is used to
32 derive the radar observables of the rimed particles.

1 **3 Case Study Analysis**

2 **3.1 Synoptic Situation in Hyytiälä on February 21, 2014**

3 On February 21, 2014 a low-pressure system situated between Iceland and the British Isles
4 influenced the weather in most parts of Europe. Several surface fronts associated with the
5 weather system moved to the East/Northeast. The most prominent one was a partially
6 occluded front that crossed Western Europe and reached the stage of a fully developed
7 occlusion further East. Due to increased vertical mixing, it had the characteristics of a warm
8 occlusion near the surface with rising temperatures behind the passage of the front. The
9 advection of warm air is indicated by a veering vertical wind profile in the radio sounding
10 launched at 23.2 UTC as shown in Figure 1. With warm air sliding over cold air, the first
11 clouds associated with this system were found at higher levels (around 8 km).

12 Prior to the arrival of the warm occlusion in Hyytiälä, multiple cloud layers were present
13 which are obvious in the radio sonde launched at 23.2 UTC. The cloud-base of the frontal
14 system was continuously lowering with the approaching warm occlusion. Also, the
15 reflectivity of the cloud as observed by the 35-GHz vertically-pointing cloud radar KAZR (cf.
16 Figure 2) showed tilted fall streak features above 3.5 km altitude indicating vertical wind
17 shear consistent with the radio sounding profile of horizontal wind. The temperature profile
18 shows two inversion layers; a boundary layer inversion at 0.5-0.8 km at which a shallow low-
19 level cloud had formed and a second one at 2.8-3 km where a midlevel cloud had formed as
20 illustrated in Figure 3. Cloud top temperatures of the warm occlusion front and the midlevel
21 cloud were $-50\text{ }^{\circ}\text{C}$ (at 8 km) and $-12\text{ }^{\circ}\text{C}$ (at 3.4 km), while the surface temperature was $-3\text{ }^{\circ}\text{C}$,
22 respectively. The ambient relative humidity profile shows layers of saturated/sub-saturated
23 conditions associated with the two distinct cloud layers extending from 2.3 km and higher as
24 well as from 0.2 km to 0.9 km. Subsaturated conditions leading to sublimation prevailed
25 between 0.8 km to 2.3 km as well as below 0.2 km. Thus, snowfall associated with the onset
26 of the frontal system (22.7-22.8 UTC) experienced sublimation before reaching the ground.

27 **3.2 In-situ observations**

28 During the period of interest when the snow front moved in (22.7 – 22.8 UTC), in-situ
29 observations showed low snowfall rates below (0.3 mm h^{-1}) and low total particle
30 concentrations ($< 100\text{ m}^{-3}$). PIP images were often out-of-focus; however, the structure of

1 several individual ice particles was identified: In addition to small and large oriented
2 dendrites, fast-falling roundish particles with high density – an indication of riming – were
3 observed. In the PIP 22 minute time integration interval 22.52 – 22.88 UTC, area-equivalent
4 maximum observed diameters were less than 1.5 mm (D. Moisseev, personal
5 communication). Afterwards (PIP integration time interval 22.88 – 23.06 UTC), the
6 maximum particle size increased to 3 mm and snowfall rate was still low but doubled to 0.6
7 mm h⁻¹. Later on (after 23.06 UTC) heavy snowfall of large low-density aggregates was
8 observed with the ground-based instruments and multi-frequency radar measurements as
9 discussed in Kneifel et al. (2015). During the time of interest of this study, the X-SACR was
10 not operated in vertically pointing mode.

11 **3.3 KAZR and MWR observations**

12 In Figure 2 the time-height plots of the first three moments (effective radar reflectivity factor
13 Z_e , subsequently called reflectivity; mean Doppler velocity V_d , and spectral width σ) of the
14 primary peak of the KAZR Doppler spectrum are shown. The primary peak is defined to be
15 the noise-separated peak containing the bin with maximum power spectral density (cf. Figure
16 4; Kollias et al., 2007b). This is the peak used in the standard ARM radar moments data
17 products. The ARM MicroARSCL data product (Kollias et al., 2007b) extends the reported
18 moments to skewness and kurtosis for both the primary peak, and an additional noise-
19 separated secondary peak, if one exists. The peak power densities and modal velocities of up
20 to two local maxima occurring within the primary peak are also reported. Mean and
21 maximum spectral noise power are determined using the technique described in Hildebrand
22 and Sekhon (1974). If only one hydrometeor population is present in the radar volume (cf.
23 Figure 4a), the radar Doppler spectrum is usually characterized by a single peak above noise
24 floor, which is controlled by the width of the PSD and sub-volume turbulence. In cases with
25 more than one hydrometeor class in the radar volume (e.g, liquid droplets and snow) it is
26 possible to have sufficient fall velocity separation between the two hydrometeor classes so
27 that the radar Doppler spectra is bimodal. As mentioned in Luke and Kollias (2013), strongly
28 multimodal situations can be considered to be “golden” samples as they make it easy to
29 separate the contributions of the individual hydrometeor populations to the total radar return.
30 Here, we classify all peaks which are separated by the mean noise floor into three categories
31 (liquid droplets, freshly generated ice, and (rimed) snow). Peaks are grouped into these
32 classes according to their mean Doppler velocity and spectrum width. However, in many

1 observations, the terminal fall velocity difference between two different particle size
2 distributions is not large enough to produce individual peaks separated by the mean noise
3 floor. Instead, broad merged peaks consisting of the contribution of two or more PSD occur
4 (cf. Figure 4d). This is also the reason, why only a short time period is analyzed in this study:
5 Synoptic situations in which ice particles from a „seeder“ cloud above are falling through a
6 SLW layer where they experience riming occurred at least half a dozen time during the
7 BA ECC-SNEX period but unfortunately, for all other events, only merged peaks were
8 observed in the KAZR Doppler spectra.

9 An objective way to distinguish spherical supercooled liquid droplets from freshly generated
10 non-spherical ice would be the use of spectral linear depolarization ratio (LDR).
11 Unfortunately though, no KAZR crosspolarization channel data was gathered during BA ECC-
12 SNEX and thus no LDR could be determined. However, microphysical modeling sensitivity
13 tests (not shown) showed that unrealistically high ice particle number concentrations - on the
14 order of a few hundreds to a few thousands per liter for ice particle sizes of a few hundred
15 microns - would be required to produce a peak of about -20 to -15 dBZ at 0.15 m s^{-1} . For this
16 regime of sizes, the observed ice particle number concentration reported in the literature is
17 below 10 per liter (e.g., Zhang et al. 2014, Lloyd et al. 2015). We can thus assume that power
18 spectrum peaks at small velocities are due to liquid droplets and not ice particles.

19 In this study, the focus is on the period when the upper level snow band moves in, roughly
20 from 22.4 UTC when first detected by the KAZR to the end of the first snow shower at
21 around 22.77 UTC. In the KAZR reflectivity field (cf. Figure 2a) the onset of snowfall is
22 clearly visible by the high Z_e ($> -5 \text{ dBZ}$) area extending from a fall streak feature at a height
23 of 8 km downwards. The Z_e structure is strongly tilted above 3.5 km due to horizontal wind
24 shear. The decrease in Z_e at 22.7-22.77 UTC below 1 km can likely be attributed to
25 sublimation in an ice-subsaturated layer (cf. humidity profile in Figure 1).

26 Before the arrival of the snow band in Hyytiälä, a liquid-topped mixed-phase cloud with
27 cloud top at 3.4 km is observed (see Figure 3). Its roughly 500 m thick SLW layer can clearly
28 be distinguished by mean Doppler velocities of around 0.0 m s^{-1} in Figure 2b, as well as very
29 narrow spectrum width values below 0.08 m s^{-1} and low reflectivity values ($Z_e < -15 \text{ dBZ}$). A
30 typical KAZR Doppler spectrum example within this liquid layer is shown in Figure 4a.
31 Within this SLW layer new ice formation took place at about $-12 \text{ }^\circ\text{C}$; subsequently, the ice
32 sedimented and grew in size as reflected by a gradual increase of Z_e (22.4-22.69 UTC)

1 between 2.9-0.9 km, as well as an increase of V_d , and σ . A Doppler spectrum example of this
2 freshly generated ice mode is shown in Figure 4c. The rapid increase of Z_e below 0.9 km
3 prior to the snowfall indicates ice particle riming which is confirmed by the ground-based in-
4 situ observations of rimed ice crystals during that time period. The riming most likely
5 occurred in another SLW layer at the lower inversion at 0.7-0.9 km, characterized by mostly
6 not noise-floor separated liquid peaks in the Doppler spectra (not shown). Strong surface
7 turbulence below 0.8 km is obvious in highly variable V_d and high σ . The strong turbulence
8 in the surface layer led to a broadening of the Doppler spectra peaks which resulted in broad
9 merged peaks (not separated by the noise-floor) and thus hampered the application of our
10 microphysical retrievals for that SLW layer.

11 When the snow band starts falling through the SLW layer, the primary peak moments are not
12 sufficient to capture the radar view of the microphysics. The coincidence of liquid and ice
13 particle size distributions within the KAZR sampling volume leads to multi-modal KAZR
14 radar Doppler spectra due to the terminal fall velocity difference between the liquid droplets
15 and the falling snow (Figure 4b). As obvious in Figure 4b, the snow falling through the SLW
16 layer has a higher dynamic range than the liquid mode and is thus classified as the principle
17 peak, leading to a sudden change of V_d values in Figure 2b after 22.69 UTC and below 3.4
18 km. Similarly, the sudden high values of σ at 22.68-22.7UTC and below 2 km can be
19 explained by a merged peak of the freshly generated ice and the snow which are no longer
20 separated by the mean noise floor, as illustrated in Figure 4d. The strong increase of V_d (on
21 the order of 0.5 m s^{-1} , cf. Figure 2b) within the SLW layer indicates riming of particles, as
22 also seen in the in-situ measurements (cf. Section 3.2).

23 Observations supporting the presence of riming when the snow fall streak intercepts the SLW
24 layer are provided in the temporal evolution of the MWR liquid water path (LWP) shown in
25 Figure 5b. While the LWP varies between $320\text{-}400 \text{ g m}^{-2}$ before 22.69 UTC when only the
26 liquid-topped mixed-phase cloud is present, it rapidly decreases from 400 g m^{-2} to 250 g m^{-2}
27 within 3 min (22.69-22.74 UTC). The observed reduction in the LWP can partly be attributed
28 to the capturing of SLW droplets by the falling snowflakes, leading to rimed particles with
29 high density and fast fall velocities.

30 It should be noted that the vertically integrated value of LWP cannot be entirely attributed to
31 the layer of SLW at 2.9-3.4 km. Unfortunately, the entire vertical distribution of SLW cannot
32 be reconstructed because the lidars are already extinguished by a thin SLW layer at 0.2-0.4



1 km. However, two more thin SLW layers were detected by KAZR Doppler spectrum analysis
2 at 0.8-0.9 km and 1.5-1.7 km (see Figure 3). These layers however were intermittent and
3 coincide in time with the periods of highest values of LWP around 22.6-22.65 UTC and 22.7
4 UTC. When these intermittent layers were observed, the LWP increased by 60-80 g m^{-2} ; we
5 use this values as the estimate of their combined contribution to the total LWP. The SLW
6 layer that had formed at the lower temperature inversion at 0.8 km is a remnant signature of
7 the long-lived thin mixed-phase cloud with cloud top at just below 0.9 km that was present
8 since before 14 UTC.

9 Based on the liquid peak reflectivity (cf. Figure 5c) of the SLW layer at 2.9-3.4 km the mean
10 liquid water content of this layer of 80-90 g m^{-2} (averaged between the two slanted paths in
11 Figure 5) was estimated (cf. Figure 10).

12

13 3.3.1 Fall streak tracking

14 When vertical wind shear is observed, the analysis of vertical profiles is not sufficient to
15 correctly trace the paths of evolution of hydrometeor populations, as already stated in
16 Marshall (1953). Instead, it is necessary to follow the falling hydrometeor populations along
17 slanted fall streaks if we want to estimate their microphysical evolution from the particle
18 generating level (z_{gen}) to height z . As illustrated in Hogan and Kew (2005), the slanted fall
19 streak patterns can be simulated in the radar time-height observation space by taking the mean
20 Doppler velocity V_d as average fall velocity of the particle population and using the horizontal
21 wind profile $u(z)$, e.g. from radio soundings closest in time for advection:

$$22 \quad \frac{dz}{dt_{\text{rad}}} = (V_d(z) u(z_{\text{gen}})) / (u(z) - u(z_{\text{gen}})). \quad (1)$$

23 Here, $u(z_{\text{gen}})$ is the horizontal wind velocity at the particle generating level z_{gen} which is
24 assumed to be close to cloud top ($z_{\text{gen}} = 6 \text{ km}$) and t_{rad} is the radar time (x-axis in time-height
25 plots in vertically pointing mode). It is important to note that t_{rad} is usually different from the
26 sedimentation time of the particle population and that fall streaks are not identical to particle
27 trajectories but are a result of overlapping trajectories as they are advected over the radar
28 (Bohren and Fraser, 1992). As in Hogan and Kew (2005), we use the finite difference
29 equivalent of Equation (1) and work our way downward level-by-level considering the
30 displacement at level z in the determination of displacement at level $z-1$. We use this
31 technique to determine a fall streak for each radar time step. Subsequently, all averaged


1 profiles (means and standard deviations) are determined for the fall streaks spanned between
2 the two black slanted fall streaks in Figure 5. Our focus of interest will be the uppermost SLW
3 layer at the height range 2.9-3.4 km.

4 The closest radio sounding in time was launched at 23.2 UTC, approximately 45 min after the
5 snow front was first observed with the KAZR. Ideally, horizontal wind profiles at the time of
6 interest should be used; however, overall agreement of the slope of the black simulated fall
7 streaks in Figure 5 with the slope of the reflectivity features (above 2.9 km) seems to confirm
8 that the horizontal wind field did not change considerably within these 45 min. Below 3 km
9 the slope of the black simulated fall streaks does not fit well with the Z_e feature, which is
10 tilted in the opposite direction compared to the fall streak above. Additional fall streak
11 simulations (depicted in blue in Figure 3) show that this backward tilted slope is well
12 matched by simulations if we assume $z_{gen} = 2.9-3$ km and the mean Doppler velocity of the
13 rimed snow mode in Equation (1). The lower-level Z_e feature thus corresponds to the fall
14 streaks of those snow particles that experience riming in the SLW layer and indicates – as
15 expected - the SLW layer as the correct generating level for the rimed particle population.

16

17 **3.3.2 Doppler spectrum evolution**

18 Figure 6 illustrates the necessity of tracking particle populations along slanted paths in
19 conditions of wind shear by contrasting the KAZR Doppler spectrum evolution along the
20 vertical line at 22.69 UTC in Figure 6a versus along the estimated slanted fall streak in Figure
21 6d. Clearly, the vertical range spectrogram in Figure 6b shows some non-microphysical
22 features - such as the discontinuity of spectral reflectivity at around 4 km - caused by taking a
23 vertical profile when a slanted one is more representative. The range spectrogram along the
24 slanted path in Figure 6e tells a much more consistent microphysical evolution story of the
25 particles when the frontal system moves in: Above the 3.4 km height, snow with a mean
26 Doppler velocity of around 1 m s^{-1} is observed. A few shallow layers of increased turbulence
27 as well as thicker layers of up-/downdrafts result in a shift of the entire Doppler spectrum to
28 more positive/negative values, respectively. At 2.9-3.4 km, the layer of SLW with V_d
29 fluctuating around 0 m s^{-1} is obvious in Figure 6e as well as in the time spectrogram in Figure
30 6c. Due to their small size, the terminal fall velocities of SLW droplets is negligible, so their
31 V_d (also illustrated in Figure 5d) can be used as an air motion tracer as done in previous
32 studies (e.g., Shupe et al., 2004; Rambukkange et al., 2011). The V_d offset of the liquid mode

1 from 0 m s^{-1} gives us an estimate of radar volume mean vertical air motion. Thus, in the SLW
2 layer, the fall velocity of ice particles can be corrected for vertical air motion as illustrated in
3 Figure 6f. In Figure  the same air-motion correction is applied in regions with SLW.
4 Again, the increase of V_d of the snow falling from higher layers into the SLW layers indicates
5 riming. Also, the new ice mode generated in the SLW layer is revealed, as its V_d quickly
6 increases from a few ten cm s^{-1} to 0.8 m s^{-1} between 2.8 km and 2 km.

7

8 **3.3.3 Profiles and PDF of radar moments**

9 Although the given observations cannot fully disentangle microphysical and dynamical
10 effects, a fairly consistent picture of the evolution of the present hydrometeor populations can
11 be formed (cf. Figure 6). The means and standard deviations of Z_e , V_d , and σ of slanted
12 profiles of the three hydrometeor populations (SLW, ice generated in the SLW layer, and
13 frontal snow) are shown in Figure 7. The average profile and standard deviations are based on
14 all 110 simulated fall streaks encompassed by the two dashed lines in Figure 5. Since in the
15 Rayleigh scattering regime radar reflectivity is proportional to the number of particles N and
16 to D^6 (with D being the diameter of the droplets), the reflectivity of the SLW droplets is very
17 low (-18 to -22 dBZ on average, cf. Figure 7a). The reflectivity of the ice generated in the
18 SLW layer increases between 3.2 km and 2 km from -16 to -10 dBZ because the ice particles
19 grow due to water vapor deposition and aggregation. At the same time, the V_d of the ice mode
20 increases from 0.2 to 0.7 m s^{-1} (cf. Figure 7b). The absence of an air motion tracer outside of
21 the SLW layer does not allow for vertical air motion contributions to the observed V_d to be
22 accounted for. As a result, the observed V_d is not equal to the terminal fall velocity of the ice
23 particles but could be higher (in updrafts) or lower (in downdrafts). Due to its large particles,
24 the greatest contribution to the total radar return is of course given by the frontal snow mode.
25 The total increase of mean Z_e of this mode from 6 km to 2 km is 12 dB; however, from 6 km
26 to the top of the SLW layer, the mean Z_e - and V_d increase of the snow mode is only moderate
27 (5 dB and 0.15 m s^{-1}), suggesting only moderate growth of the snow particles, which is likely
28 due to water vapor deposition in higher parts of the cloud. V_d of the frontal snow mode
29 increases from 1 m s^{-1} to 1.65 m s^{-1} between 3.2 km and 2 km. Spectrum width profiles in
30 Figure 7c show several thin turbulent layers in which the standard deviation of σ is high. σ
31 of the SLW mode is smallest (0.05 - 0.07 m s^{-1}) while σ of the new ice and frontal snow mode

1 are on the order of 0.1-0.2 m s⁻¹. As previously mentioned, larger σ can be caused by a
2 superposition of broad PSD and sub-volume turbulence.

3 Probability density functions (PDF) of the moments of the three hydrometeor populations are
4 shown in Figure 8. Only data points from below the SLW layer top (3.4 km) down to the
5 surface and between the two black dashed fall streaks in Figure 5 are considered. For Z_e and
6 V_d there is very little overlap in the PDF of each hydrometeor population. In uniform beam-
7 filling conditions and sub-volume turbulence, the Doppler spectrum of a cloud droplet PSD is
8 symmetrical and near Gaussian, resulting in zero skewness (Kollias et al., 2011). In contrast,
9 the PDF of the observed SLW mode is skewed towards negative values, indicating the
10 presence of supercooled drizzle at around -12 °C. The presence of drizzle is also indicated in
11 Figure 9 where the joint PDF of skewness and Z_e are shown. While liquid mode skewness
12 fluctuates around zero for Z_e below -20 dBZ, it becomes increasingly negative at higher Z_e
13 values of -20 to -16 dBZ. The existence of drizzle in SLW layers at the temperature range of -
14 5 to -12 °C was also found by Verlinde et al. (2013) for Arctic multi-layered mixed-phase
15 clouds; the skewness-reflectivity signature as shown in Figure 9 is qualitatively similar to the
16 signatures observed in warm drizzle clouds (Luke and Kollias, 2013).

17

18 **3.4 1D microphysical bin modeling**

19 The comprehensive observations were used as input to a one-dimensional (1D) bin-
20 microphysical model. The model is based on the warm rain model first presented in Szyrmer
21 et al. (2005) and modified to include the ice processes of deposition, aggregation, and riming.
22 The leading question is whether it is possible to reproduce the evolution of the observed radar
23 moments of the rimed snow mode in the SLW layer using radar forward modeling of the
24 microphysical model output. Forcing the model output to agree with the observations could
25 help us to evaluate different proposed riming efficiency schemes that have been implemented
26 as options in the model (see Appendix A). The microphysical modeling was limited to the
27 SLW layer between 2.9-3.4 km and focused on the evolution of the moments (Z_e , V_d , σ) of
28 the frontal snow mode.

29

1 **3.4.1 Model description**

2 The one-dimensional (1D) steady-state model simulates the height evolution of the bin-
3 resolved snow PSD introduced at the model's uppermost level (which in this study is the top
4 of the SLW layer at 3.4 km). The model input includes the vertical profiles of temperature,
5 pressure, relative humidity (taken from the radio sounding, cf. Figure 1) and vertical air
6 motion (derived from the SLW mode V_d ; cf. Figure 10). The cloud droplet PSD at each level
7 within the SLW layer is calculated from the liquid peak reflectivity profile assuming that the
8 cloud droplets follow a lognormal size distribution with a height-independent prescribed
9 droplet number concentration and dispersion parameter (as also taken for calculations of
10 liquid water content (LWC) in Figure 10).

11 The evolution of the snow mode along the fall streak is simulated by explicitly calculating the
12 contribution to particle growth by the microphysical processes of water vapor diffusion,
13 aggregation, and riming. In the model setup used in the presented simulations, the water vapor
14 deposition and aggregation processes conserve the area ratio, aspect ratio, and mass-size
15 relationships. The deposition calculations use the results of Field et al. (2008) to describe the
16 capacitance, except for the smaller particles for which the electrostatic capacitance
17 approximation for thin plates is assumed, and the ventilation factor proposed by Hall and
18 Pruppacher (1976) is adapted. The value of aggregation efficiency is set to 0.2. Details of the
19 riming parameterizations are presented in Appendix A.

20 The frontal snow PSD is introduced at the uppermost model level in a functional form of a
21 generalized gamma function with melted diameter representing particle size. The values of the
22 two shape parameters are taken from Delanoë et al. (2005) for the form that is most consistent
23 with the observations. The initial mass-size relation of a power-law form with exponent 2 and
24 prefactor 0.0012 g cm^{-2} is used. The rimed fraction of the snow particles at the model's
25 uppermost level is set to 0. The initial area ratio is calculated from the empirical relation
26 between area ratio and particle density in Heymsfield et al. (2002) for side plane aggregates
27 (particle density = $0.18 \text{ area ratio}^{1.5}$).

28

29 **3.4.2 Comparison of model output and observations**

30 In Figure 11, KAZR-observed profiles of snow mode radar moments and the modeled
31 moments are compared. Backscattering calculations are performed using Mie spheres with

1 non-uniform mass distribution (Fabry and Szyrmer, 1999). The terminal velocity of snow
2 particles - unrimed at the top and partially rimed below - is calculated based on the method
3 proposed by Heymsfield and Westbrook (2010). However, two different possibilities for the
4 evolution of the area ratio with riming are implemented. In the first approach (denoted by “1”
5 in Figure 11) for a given particle maximum diameter, the increase of the area ratio from the
6 initial value is parameterized as a function of the aspect ratio modified by riming. In the
7 second approach (denoted by “2” in Figure 11) the increase of the area ratio is smaller and is
8 obtained assuming that the initial relation between area ratio and particle density is
9 maintained, i.e., the increase of the area ratio results from the increase of mass only.

10 In addition to the sensitivity of the terminal velocity of the rimed snow particles to the
11 evolution of the area ratio relationship, the sensitivity of the modeled radar moments of the
12 rimed snow mode to different riming efficiencies is evaluated. Specifically, in one model
13 realization, the riming efficiency parameterizations from Hall (1980) and Cober and List
14 (1993) for small and large ice particles are used, respectively. Model results using this
15 particular parameterization are denoted as “Hall Cober” in Figure 11. In another model
16 realization, the riming efficiency parameterization proposed by Lohmann (2004) assuming
17 plates for small D and aggregates for large D is adapted. Model results using this approach are
18 denoted as “Lohmann” in Figure 11.

19 Two effects can explain some of the discrepancies between the observed and modeled
20 profiles. First, dynamical effects such as turbulence are not included in the radar forward
21 model that estimates the radar Doppler spectrum width profile. Thus, the forward simulation
22 cannot reproduce the spectrum broadening of the rimed mode below 3 km (cf. Figure 11c).
23 Second, there are uncertainties in the exact slope of the fall streaks due to the fact that the
24 horizontal wind profile from the sounding is not taken at the time of interest but 45 min later.
25 Overall, the evolution of the rimed snow mode moments in the SLW layer can be reproduced
26 by the model (within the standard deviations). In Figure 11a, the observed and modeled Z_e
27 profile shows an increase from SLW top to about 3.15 km which is attributed to an increase of
28 mass of the snow when the liquid droplets attach to the snowflakes. Below 3.15 km, the
29 dZ_e/dz is slightly reduced because even with the increase of mass, the increase of backscatter
30 is very low and at the same time there is a reduction of particle number concentration. Except
31 for the aforementioned influence of turbulence on spectrum width, the model reproduces the
32 mean Doppler velocity and spectrum width profiles well. With the increase in mass and

1 density during the riming process, particle fall velocities of the rimed snow steadily increase
2 from top to bottom of the SLW layer from about 0.95 m s^{-1} to 1.2 m s^{-1} . The model is also
3 capable of reproducing the small decrease in the spectrum width due to differences in the
4 riming efficiency of small and large particles leading to a small reduction in the spread of the
5 rimed particles' fall velocities.

6

7 **4 Summary and Conclusions**

8 For a frontal winter snowfall event observed at ~~the University of Helsinki's~~ Hyytiälä field
9 ~~station~~ in Finland during the BAECC-SNEX field experiment in 2014, we show that by
10 disentangling the contributions of the different hydrometeor populations to total vertically-
11 pointing cloud radar returns, it is possible to follow the microphysical evolution of the present
12 cloud- and precipitation particles. For that purpose, we work with the entire radar Doppler
13 spectrum instead of only considering cloud radar moments, which are integrated parameters
14 of the spectrum.

15 The analysis presented here focuses on a band of snow that falls through a SLW layer, where
16 it experiences riming and where new ice particle formation also takes place. A detailed
17 analysis of the vertical evolution of radar moments of the frontal snow, SLW droplets, and
18 freshly generated ice in terms of their evolution of radar moments (Ze , V_d , σ) is presented.

19 The extensive analysis accounts for the vertical shear of the horizontal wind and the tracking
20 of particle populations is performed along slanted paths instead of vertical profiles.
21 Furthermore, the multiple noise-floor-separated modes of the radar Doppler spectra are
22 analyzed and the moments of the SLW, ice, and rimed snow determined separately. From the
23 V_d and Ze of the SLW droplets we estimate vertical air motion and LWC within the SLW
24 layer, respectively.

25 The observations were used to set up and evaluate a 1D steady-state bin microphysical model
26 that accounts for the processes of deposition, aggregation and riming. In particular, the
27 profiles of vertical air motion and SLW content, along with thermodynamic variables from
28 the nearest sounding were used to set up the steady-state conditions within the SLW layer.
29 The radar moments at the top of the SLW layer were used to initiate the unrimed snow PSD.
30 The radar Doppler spectrum analysis enabled isolation of the radar moments of the rimed
31 snow within the SLW layer.

1 The microphysical model output (PSD of rimed snow particles) is used as input to a simple
2 radar forward model that reproduces the three radar Doppler moments. The scattering model
3 used to estimate the backscattering cross-section of the snow particles is based on a two-layer
4 spherical model introduced by Fabry and Szyrmer (1999). The uncertainties associated with
5 this backscattering computation have been discussed in Szyrmer et al. (2012). While there are
6 several different methods to prescribe the scattering calculations of snow particles (e.g.,
7 Tyynelä et al., 2013; Hogan and Westbrook, 2014), the height evolution of the radar moments
8 of the rimed snow mode is not very sensitive to the scattering method used. Also, the very
9 small particle sizes observed at the surface indicate that larger snowflakes which produce
10 more complex Mie scattering, as for example seen later on February 21, 2014 (Kneifel et al.,
11 2015), were not present during the period of interest.

12 Assuming a generalized gamma function with melted diameter representing particle size and
13 using two different methods to estimate the terminal fall velocity of the rimed particles as
14 well as two different methods to estimate the riming efficiency of the snow particles, we were
15 able to reproduce the observed profiles (Figure 11). Modeling was also performed assuming
16 an exponential snow PSD form (not shown). For this PSD form it is found that the modeled
17 profiles of the rimed snow Ze and V_d are similar to those modeled using the generalized
18 gamma function PSD. However, for the exponential PSD assumption the modeled radar
19 Doppler spectrum width profile had much less sensitivity to riming efficiency
20 parameterization or choice of area ratio increase.

21 The case presented here was carefully selected to represent a scenario where riming is the
22 dominant process that modulates the snow radar moments (riming fingerprinting). The
23 simulations indicate that for fixed parameterizations for deposition and aggregation, the
24 number of combinations of the riming and velocity parameterizations resulting in profiles that
25 are comparable to the observations is rather limited. However, changing any of these factors
26 may lead to different results. In other words, the effect of choosing different riming efficiency
27 parameterizations is on the same order as choosing different options for area ratio increase,
28 and thus fall velocity calculations, as indicated by a equidistant spread of modeled profiles of
29 Ze , and V_d in Figure 11.

30 Thus, no clear conclusions can be made as to which riming parameterization is more
31 appropriate in this particular case due to large sensitivity of the model simulations to several
32 factors other than those we assume, because of a lack of additional observational constraints.

1 The profiling radar observations provide a height evolution perspective, which is a critical
2 constraint for fingerprinting studies that aim to “isolate” a vertical layer where a particular
3 microphysical process dominates particle growth. However, as it is evident from the
4 sensitivity of the simulations, in-situ observations giving exact velocity-size relations, particle
5 size- and density estimates, and information about particle shape are needed to further limit
6 the choices in the model setup leading to true advancement in our understanding of riming
7 and other ice microphysical processes. While extensive ground-based in-situ data was
8 gathered during BAECC-SNEX, the period of interest was characterized by low precipitation
9 rates which hampered the ground-based microphysical retrievals. Furthermore, measurements
10 of the LWC and ice PSD at the top of the SLW layer would have been key parameters to
11 constrain the model runs. This points to the future need for coordinated (aircraft-based) in-situ
12 observations combined with profiling and scanning radar observations. The DOE ARM
13 Program and the ARM Aerial Facility (AAF) are well positioned to conduct such targeted
14 observational-modeling studies in the future.

15 The present riming case study was selected from the extensive BAECC campaign data set due
16 to its nearly ideal situation in which the fall velocity separation of the different hydrometeor
17 classes is strong enough to produce individual peaks separated by the mean noise floor in the
18 cloud radar Doppler spectra. During other riming cases, cloud radar Doppler spectra
19 multimodalities were observed, however, these peaks were usually merged. These cases are
20 more complicated to disentangle, most likely due to a more complex mixture of dominant ice
21 growth processes (water vapor deposition, aggregation, and riming) all played an important
22 role. Fingerprinting studies of *one* particular microphysical growth process - such as riming -
23 requires more simple microphysical situations as we think is the case in the presented study.

24 In situations with merged broad peaks it is very challenging to define *objective* peak
25 separation criteria and to thus disentangle the relative contributions of different hydrometeor
26 populations to the total radar returns. In a previous mixed-phase cloud radar Doppler spectra
27 study (Shupe et al., 2004) empirical “peak-picking” criteria were developed by manual
28 inspection of the peak-picking results. There, it was emphasized that the criteria depend on
29 the observed mixed-phase cloud cases and cloud radar sampling parameters (such as temporal
30 resolution and number of FFT points). The development of robust cloud radar Doppler spectra
31 peak separation criteria in mixed-phase clouds is the topic of future studies, the dataset
32 gathered during the BAECC campaign offers great potential for such studies.

33

1

2 **Appendix A: Parameterizations of Riming Efficiency and Physics**

3 The increase in mass of an individual snow particle via the riming process is calculated with
4 the stochastic collection equation. Different options for the calculations of the riming
5 efficiency, the evolution of area ratio and aspect ratio via the riming process, and the related
6 increase in particle fall velocity are included in the model. Size-dependent riming efficiency
7 parameterizations proposed in the literature and introduced in the model can be separated into
8 two groups. The first group describes the efficiency of smaller pristine crystals, mainly at the
9 first stage of riming, based on numerical simulations of Pitter and Pruppacher (1974); Pitter
10 (1977); and Wang and Ji (2000), and proposed by Hall (1980); Young (1993); Geresdi
11 (1998); and Lohman (2004) taken from Mitchell (1990). The second group of
12 parameterizations introduced in the model are more suitable for larger more spherical
13 particles: Included are the parameterization developed for the accretional growth of raindrops
14 (Beard and Grover, 1974), a parameterization derived for graupel (Cober and List, 1993) and
15 the one proposed by Lohmann (2004) for aggregates based on the experimental results of Lew
16 et al. (1986). Some examples of the dependence of the riming efficiency on the snow particle
17 size calculated for different cloud droplet diameters are shown ~~in~~ in the supporting
18 information of Leinonen and Szyrmer (2015).

19 Different approaches to describing the physics of riming result in different descriptions of the
20 changes of the properties of particles undergoing riming growth. Aggregates and branched
21 particles appear to grow by “filling in”, resulting in an increase of particle effective density
22 while the major dimension does not change. Mainly, the minor dimension is expected to
23 increase. But when the particle effective density is large enough, associated with a quasi-
24 spherical shape, the filling process has to be replaced by an increase of both dimensions with
25 the aspect ratio maintained (e.g., Morrison and Grabowski, 2010). Different options of change
26 of particle aspect ratio accompanying the growth by riming in the model use the rime density
27 calculated from empirical formulas (Macklin, 1962; Pflaum and Pruppacher, 1979;
28 Heymsfield and Pflaum, 1985). The following options for the evolution of the area ratio for a
29 given increase of the rimed fraction can be selected in the model: i) using one of the empirical
30 relations of mass-density-area ratio (by choosing an appropriate relation from the Table in
31 Szyrmer et al., 2012 or others), or ii) obtained by interpolation based on rimed fraction
32 between the values associated with the unrimed particle and graupel (as in Lin and Colle,

1 2011), or iii) calculated from the assumed relation of particle geometry between area ratio and
2 aspect ratio (e.g. Avramov et al., 2011; Jensen and Harrington, 2015).

3

4 **Data availability**

5 All data used in this study is publicly accessible at the ARM Data Archive:
6 www.archive.arm.gov.

7

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16

17 **References**

- 18 Avramov, A., et al., 2011: Toward ice formation closure in Arctic mixed-phase boundary
19 layer clouds during ISDAC, *J. Geophys. Res.*, 116, D00T08, doi:10.1029/2011JD015910.
- 20 Barrett A., et al., 2010: Improving mixed-phase cloud representation in weather and climate
21 models, Poster contribution at AMS Clouds and Radiation Conference, Portland, OR.
- 22 Baschek, B., R. Schefold, and E. Barthazy, 2004: How do updrafts and embedded convection
23 influence riming?, *Proceedings of ERAD*, 261-267.
- 24 Beard, K. V., and S. N. Grover, 1974: Numerical collision efficiencies for small raindrops
25 colliding with micron size particles. *J. Atmos. Sci.*, 31, 543–550.
- 26 Bohren C.F. and A. B. Fraser, 1992: Fall streaks: Parabolic trajectories with a twist. *American*
27 *Journal of Physics*, 60 (1030), doi: 10.1119/1.16982.

28

1 Borys, R. D., D. H. Lowenthal, S. A. Cohn, and W. O. J. Brown, 2003: Mountaintop and
2 radar measurements of anthropogenic aerosol effects on snow growth and snowfall rate,
3 *Geophys. Res. Lett.*, 30(10), 1538, doi:10.1029/2002GL016855.

4 Cadeddu, M. P., J. C. Liljegren, and D. D. Turner, 2013: The Atmospheric radiation
5 measurement (ARM) program network of microwave radiometers: instrumentation, data, and
6 retrievals, *Atmos. Meas. Tech.*, 6 (9), 2359-2372, doi:10.5194/amt-6-2359-2013.

7 Cantrell, W., and A. Heymsfield, 2005: Production of Ice in Tropospheric Clouds, *Bull.*
8 *Amer. Meteor. Soc.*, 86, 795-807.

9 Cober, S. and R. List, 1993: Measurements of the Heat and Mass Transfer Parameters
10 Characterizing Conical Graupel Growth. *J. Atmos. Sci.*, 50, 1591–1609.

11 Delanoë, J., A. Protat, J. Testud, D. Bouniol, A. J. Heymsfield, A. Bansemer, P. R. A. Brown,
12 and R. M. Forbes, 2005: Statistical properties of the normalized ice particle size distribution,
13 *J. Geophys. Res.*, 110, D10201, doi:10.1029/2004JD005405.

14 Fabry, F., and Szyrmer, W., 1999: Modeling of the melting layer. Part II: Electromagnetics, *J.*
15 *Atmos. Sci.*, 56: 3593-3600.

16 Field, P.R., A.J. Heymsfield, A. Bansemer, C. H. Twohy, 2008: Determination of the
17 combined ventilation factor and capacitance for ice crystal aggregates from airborne
18 observations in a tropical anvil cloud. *J. Atmos. Sci.*, 65, 376–391.

19 Gregory, D., and D. Morris, 1996: The sensitivity of climate simulations to the specification
20 of mixed phase clouds, *Climate Dyn.*, 12, 641–651.

21 Hall, W.D., and H. R. Pruppacher, 1976: The survival of ice particles falling from
22 cirrus clouds in subsaturated air. *J. Atmos. Sci.*, 33, 1995-2006.

23 Hall, W.D., 1980: A Detailed Microphysical Model Within a Two-Dimensional Dynamic
24 Framework: Model Description and Preliminary Results. *J. Atmos. Sci.*, 37, 2486–2507.

25 Hari, P., and M. Kulmala, 2005: Station for Measuring Ecosystem–Atmosphere Relations
26 (SMEAR II). *Boreal Env. Res.*, 10, 315–322.

27 Heymsfield, A., and J. Pflaum, 1985: A quantitative assessment of the accuracy of techniques
28 for calculating graupel growth. *J. Atmos. Sci.*, 42, 2264–2274.

1 Heymsfield, A., S. Lewis, A. Bansemer, J. Iaquinta, L. M. Miloshevich, M. Kajikawa, C.
2 Twohy, and M. R. Poellot, 2002: A general approach for deriving the properties of cirrus and
3 stratiform ice cloud particles. *J. Atmos. Sci.*, 59, 3–29. doi: 10.1175/1520-
4 0469(2002)059<0003:AGAFDT>2.0.CO;2

5 Heymsfield, A., and C. Westbrook, 2010: Advances in the estimation of ice particle fall
6 speeds using laboratory and field measurements. *J. Atmos. Sci.*, 67, 2469–2482.

7 Hildebrand, P. and R. Sekhon, 1974: Objective Determination of the Noise Level in Doppler
8 Spectra., *Journal of Appl. Met.*, 13, Issue 7, 808-811.

9 Hogan, R. J., and S. F. Kew, 2005: A 3D stochastic cloud model for investigating the
10 radiative properties of inhomogeneous cirrus clouds, *Q. J. R. Meteorol. Soc.*, 131 (611, A),
11 2585-2608, doi:10.1256/qj.04.144.

12 Illingworth, A. J., et al., 2007: CLOUDNET: Continuous Evaluation of Cloud Profiles in
13 Seven Operational Models Using Ground-Based Observations, *Bull. Amer. Meteor. Soc.*, 88.

14 Jensen, A., and J. Harrington, 2015: Modeling ice crystal aspect ratio evolution during riming:
15 A single-particle growth model. *J. Atmos. Sci.*, 72, 2569-2590, doi:10.1175/JAS-D-14-
16 0297.1.

17 Kanitz, T., et al, 2011: Contrasting the impact of aerosols at northern and southern
18 midlatitudes on heterogeneous ice formation, *Geophys. Res. Lett.*, 38, L17802.

19 Klein, S., et al., 2013: Are climate model simulations of clouds improving? An evaluation
20 using the ISCCP simulator, *J. Geophys. Res.*, 118, 1329–1342.

21 Kneifel, S., A. von Lerber, J. Tiira, D. Moisseev, P. Kollias, J. Leinonen, 2015: Observed
22 relations between snowfall microphysics and triple-frequency radar measurements, *J.*
23 *Geophys. Res.* (accepted).

24 Kollias, P., B. A. Albrecht, R. Lhermitte, and A. Savtchenko, 2001: Radar observations of
25 updrafts, downdrafts, and turbulence in fair-weather cumuli, *J. Atmos. Sci.*, 58, 1750–1766.

26 Kollias, P., E. E. Clothiaux, M. A. Miller, B. A. Albrecht, G. L. Stephens, and T. P.
27 Ackerman 2007a: Millimeter-wavelength radars – New frontier in atmospheric cloud and
28 precipitation research, *Bull. Am. Meteorol. Soc.*, 88, 1608–1624.

29 Kollias, P., M. A. Miller, E. P. Luke, K. L. Johnson, E. E. Clothiaux, K. P. Moran, K. B.
30 Widener, and B. A. Albrecht, 2007b: The Atmospheric Radiation Measurement Program

1 cloud profiling radars: Second-generation sampling strategies, processing and cloud data
2 products, *J. Atmos. Oceanic Technol.*, 24, 1199–1214.

3 Kollias, P., J. Remillard, E. Luke, and W. Szyrmer, 2011: Cloud radar Doppler spectra in
4 drizzling stratiform clouds: 1. Forward modeling and remote sensing applications, *J.*
5 *Geophys. Res.*, 116, D13201, doi:10.1029/2010JD015237.

6 Kollias, P., N. Bharadwaj, K. Widener, I. Jo, and K. Johnson, 2014: Scanning ARM Cloud
7 Radars. Part I: Operational Sampling Strategies, *J. Atmos. Oceanic Technol.*, 31, 569–582.

8 Lebo, Z. J., N. C. Johnson, and J. Y. Harrington, 2008: Radiative influences on ice crystal and
9 droplet growth within mixed-phase stratus clouds, *J. Geophys. Res.*, 113, D09203,
10 doi:10.1029/2007JD009262.

11 Leinonen, J. and W. Szyrmer, 2015: Radar signatures of snowflake riming: a modeling study,
12 *Earth and Space Science*, 2, 346-358, doi:10.1002/2015EA000102.

13 Lew, J. K., D. C. Montague, H. R. Pruppacher, and R. M. Rasmussen, 1986: A wind tunnel
14 investigation on the riming of snowflakes. Part II: Natural and synthetic aggregates, *J. Atmos.*
15 *Sci.*, 43(21), 2410–2417, doi:10.1175/1520-0469(1986)043<2410:AWTIOT>2.0.CO;2.

16 Lin, Y. and B. Colle, 2011: A new bulk microphysical scheme that includes riming intensity
17 and temperature-dependent ice characteristics, *Monthly Weather Review*, 139, 1013-1035.

18 Lloyd, G., T.W. Choullarton, K.N. Bower, J. Crosier, H. Jones, J.R. Dorsey, M.W. Gallagher,
19 P. Connolly, A. Kirchgaessner, and T. Lachlan-Cope, 2015: Observations and comparisons
20 of cloud microphysical properties in spring and summertime Arctic stratocumulus clouds
21 during the ACCACIA campaign, *Atmos. Chem. Phys.*, 15, 3719-3737, doi:10.5194/acp-15-
22 3719-2015.

23 Lohmann, U., 2004: Can Anthropogenic Aerosols Decrease the Snowfall Rate. *J. Atmos. Sci.*,
24 61, 2457–2468.

25 Luke, E. P., P. Kollias, and M. D. Shupe, 2010: Detection of supercooled liquid in
26 mixed-phase clouds using radar Doppler spectra, *J. Geophys. Res.*, 115, D19201,
27 doi:10.1029/2009JD012884.

28 Luke, E., and P. Kollias, 2013: Separating Cloud and Drizzle Radar Moments during
29 Precipitation Onset Using Doppler Spectra, *J. Atmos. Oceanic Technol.*, 26, 167-179, 30,
30 1656-1671, DOI: 10.1175/JTECH-D-11-00195.1.

- 1 Macklin, W.C., 1962: The density and structure of ice formed by accretion. *Quart. J. Roy.*
2 *Meteor. Soc.*, 88, 30-50.
- 3 Mitchell, D. L., 1990: Evolution of snow-size spectra predicted by the growth processes of
4 diffusion, aggregation and riming. Preprints, Conf. on Cloud Physics, San Francisco, CA,
5 *Amer. Meteor. Soc.*, 270–277.
- 6 Morrison, H., G. de Boer., G. Feingold, J. Harrington, M.D. Shupe, and K. Sulia, 2012:
7 Resilience of persistent Arctic mixed-phase clouds, *Nature Geoscience*, 5, 11–17.
- 8 Mosimann, L., 1995: An improved method for determining the degree of snow crystal riming
9 by vertical Doppler radar, *Atmos. Res.*, 37, 305-323.
- 10 Newman, A. J., P. A. Kucera, and L. F. Bliven, 2009: Presenting the Snowflake Video Imager
11 (SVI), *J. Atmos. Oceanic Technol.*, 26, 167-179.
- 12 Marshall, J.S., 1953: Precipitation trajectories and patterns, *J. Meteor.*, 10.
- 13 Petäjä, T., 2013: Science Plan Biogenic Aerosols – Effects on Clouds and Climate (BAECC),
14 US Department of Energy, Office of Science, DOE/SC-ARM-13-024.
- 15 Pflaum, J. C., and H. R. Pruppacher, 1979: A wind tunnel investigation of the growth of
16 graupel initiated from frozen drops. *J. Atmos. Sci.*, 36, 680–689.
- 17 Pitter, R. L., 1977: A reexamination of riming on thin ice plates. *J. Atmos. Sci.*, 34, 684–685.
18 doi: 10.1175/1520-0469(1977)034<0684:AROROT>2.0.CO;2
- 19 Pitter, R.L., and H.R. Pruppacher, 1974: A numerical investigation of collision efficiencies of
20 simple ice plates colliding with supercooled water drops., *J.Atmos.Sci.*, 31, 551.
- 21 Pruppacher, H. and J. Klett, 1997: *Microphysics of Clouds and Precipitation*, Kluwer
22 Academic, 954 pp.
- 23 Rambukkange, P., J. Verlinde, E. Eloranta, C.Flynn, and E. Clothiaux, 2011: Using Doppler
24 Spectra to Separate Hydrometeor Populations and Analyze Ice Precipitation in Multilayered
25 Mixed-Phase Clouds, *IEEE Geoscience and Remote Sensing Letters*, 8, 1.
- 26 Shupe, M., P. Kollias, S. Matrosov, and T. Schneider, 2004: Deriving mixed-phase cloud
27 properties from Doppler radar spectra, *J. Atmos. Ocean. Tech.*, 21, 660-670.

1 Shupe, M., J. S. Daniel, G. de Boer, E. W. Eloranta, P. Kollias, E. P. Luke, C. N. Long, D. D.
2 Turner, and J. Verlinde, 2008: A Focus On Mixed-Phase Clouds. *Bull. Amer. Meteor. Soc.*,
3 89, 1549–1562.

4 Szyrmer, W., S. Laroche, and I. Zawadzki, 2005: A microphysical bulk formulation based on
5 scaling normalization of the particle size distribution. Part I: Description. *J. Atmos. Sci.*, 62,
6 4206–4221.

7 Szyrmer, W., A. Tatarevic, and P. Kollias, 2012: Ice clouds microphysical retrieval using 94-
8 GHz Doppler radar observations: Basic relations within the retrieval framework, *J. Geophys.*
9 *Res.*, 117, D14203, doi:10.1029/2011JD016675.

10 Tyynelä, J., J. Leinonen, C. D. Westbrook, D. Moisseev, and T. Nousiainen, 2013:
11 Applicability of the rayleigh-gans approximation for scattering by snow flakes at microwave
12 frequencies in vertical incidence, *J. Geophys. Res.*, 118 (4), 1826-1839,
13 doi:10.1002/jgrd.50167.

14 Verlinde, J., M. P. Rambukkange, E. E. Clothiaux, G. M. McFarquhar, and E. W. Eloranta,
15 2013: Arctic multilayered, mixed-phase cloud processes revealed in millimeter-wave cloud
16 radar Doppler spectra, *J. Geophys. Res. Atmos.*, 118, 13,199–13,213,
17 doi:10.1002/2013JD020183.

18 Wang, P. K., and W. Ji, 2000: Collision efficiencies of ice crystals at low-intermediate
19 Reynolds numbers colliding with supercool cloud droplets: A numerical study. *J. Atmos. Sci.*,
20 57, 1001–1009.

21 Weiss, R.R. and P.V. Hobbs, 1975: The use of vertically pointing Doppler radar in cloud
22 physics and weather modification studies. *J. Appl. Meteorol.*, 14, 222-231.

23 Young, K.C., 1993: *Microphysical processes in clouds*. Oxford University Press. 427 pp.

24 Yu, G., J. Verlinde, E. E. Clothiaux, and Y.-S. Chen, 2014: Mixed- phase cloud phase
25 partitioning using millimeter wavelength cloud radar Doppler velocity spectra, *J. Geophys.*
26 *Res. Atmos.*, 119, 7556–7576, doi:10.1002/2013JD021182.

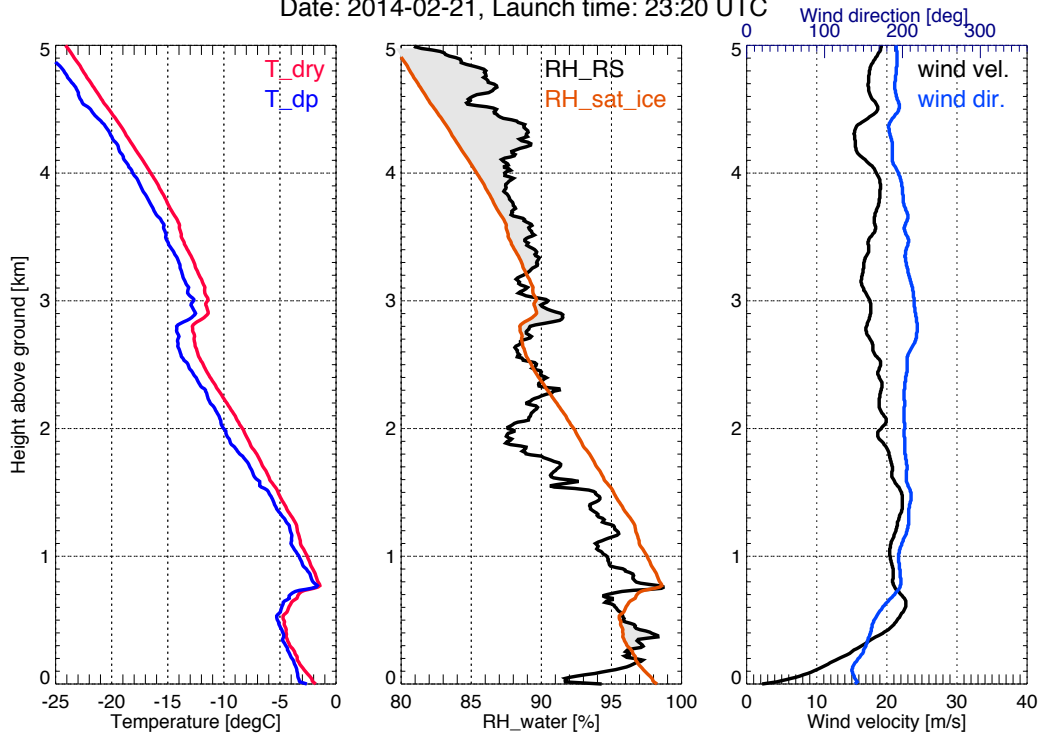
27 Zhang, D., Z. Wang, and D. Liu, 2010: A global view of midlevel liquid-layer topped
28 stratiform cloud distribution and phase partition from CALIPSO and CloudSat measurements,
29 *J. Geophys. Res.*, 115, D00H13.

1 Zhang, D., Z. Wang, A. Heymsfield, J. Fan, and T. Luo, 2014: Ice concentration retrieval in
2 stratiform mixed-phase clouds using cloud radar reflectivity measurements and 1D ice growth
3 model simulations, *J. Atmos. Sci.*, 71, 3613–3635.

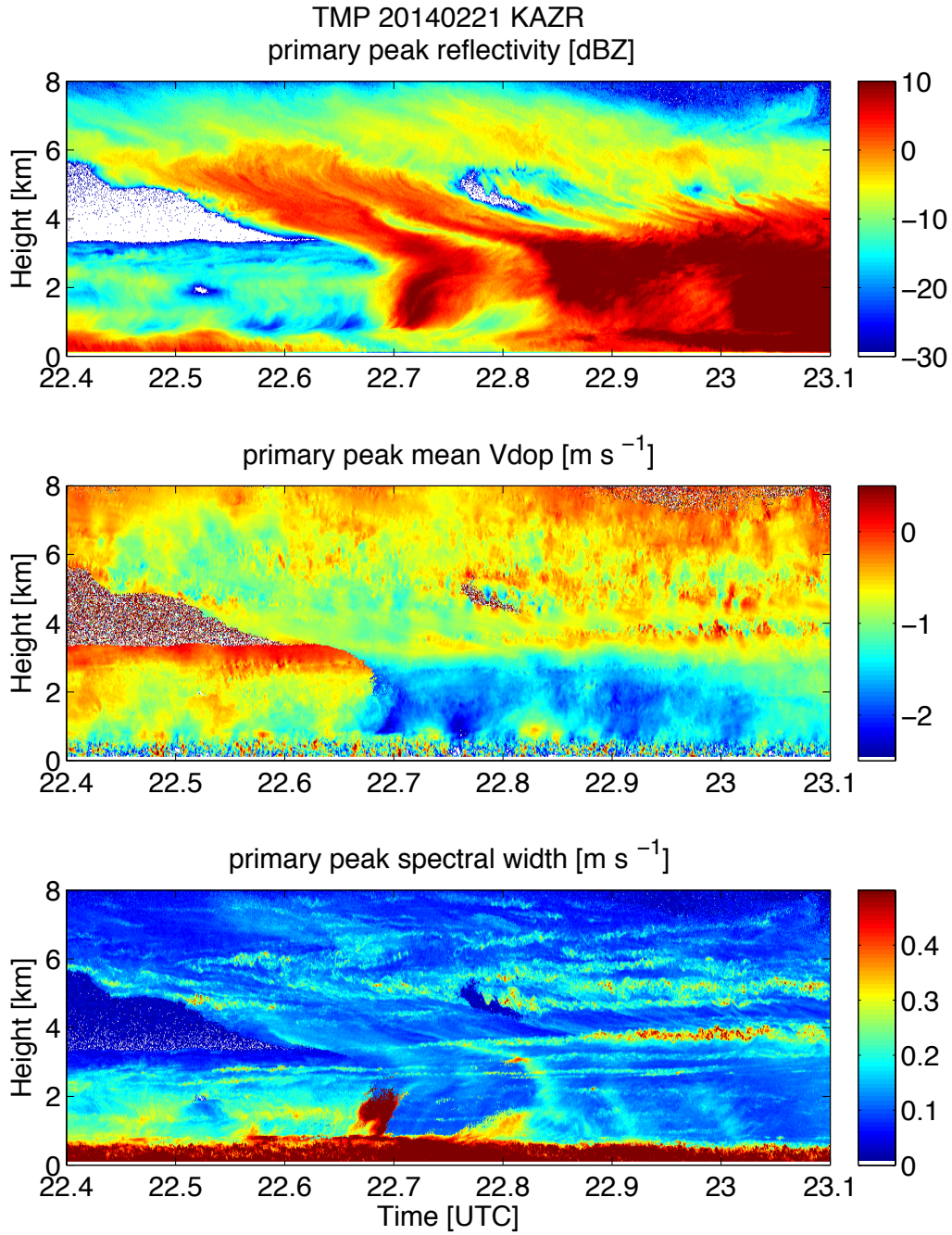
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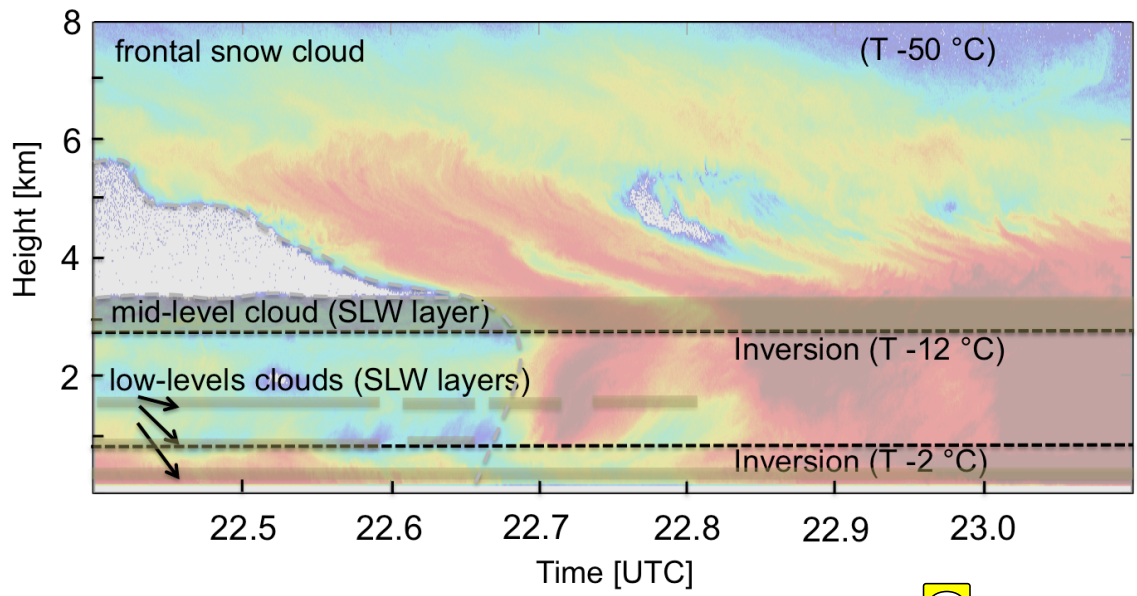


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2 Figure 1. Vertical profiles of temperature (blue, left) and dewpoint temperature (red, left),
3 relative humidity (middle, black), as well as horizontal wind speed (right, black) and wind
4 direction (right, blue) from radio sonde launched at 23:20 UTC on February 21, 2014 in
5 Hyytiälä. The red line in the middle panel refers to the humidity at which the air is saturated
6 with respect to ice, i.e., if the relative humidity is to the right of the red line, the air is
7 supersaturated w.r.t. ice (grey shading).
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Figure 2. : KAZR observed primary Doppler spectrum peak moments on February 21, 2014 in Hyytiälä. Panel a) shows reflectivity (dBZ), b) mean Doppler velocity (m s⁻¹), and c) spectrum width of the primary Doppler spectrum peak (m s⁻¹). Negative Doppler velocities indicate downward motion.



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2 **Figure 3:** Schematic diagram of the cloud layers present in Hyytiälä on 21 February 2014. Sketch is
 3 overlying the KAZR reflectivity.



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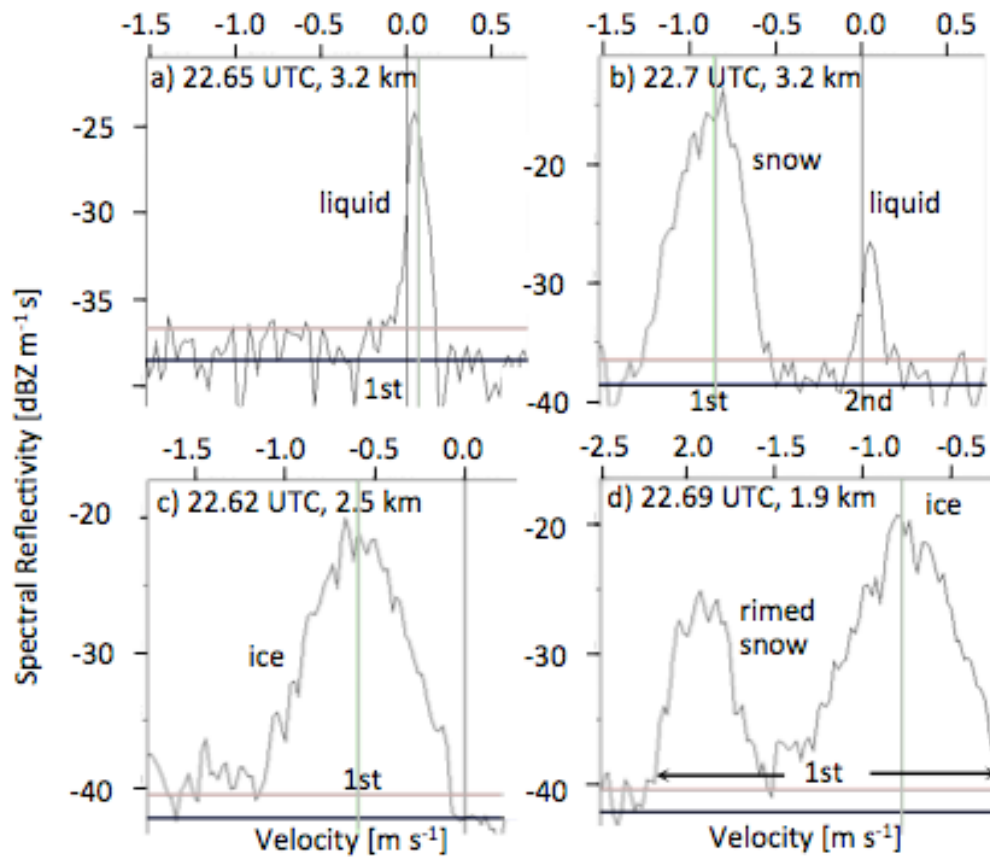
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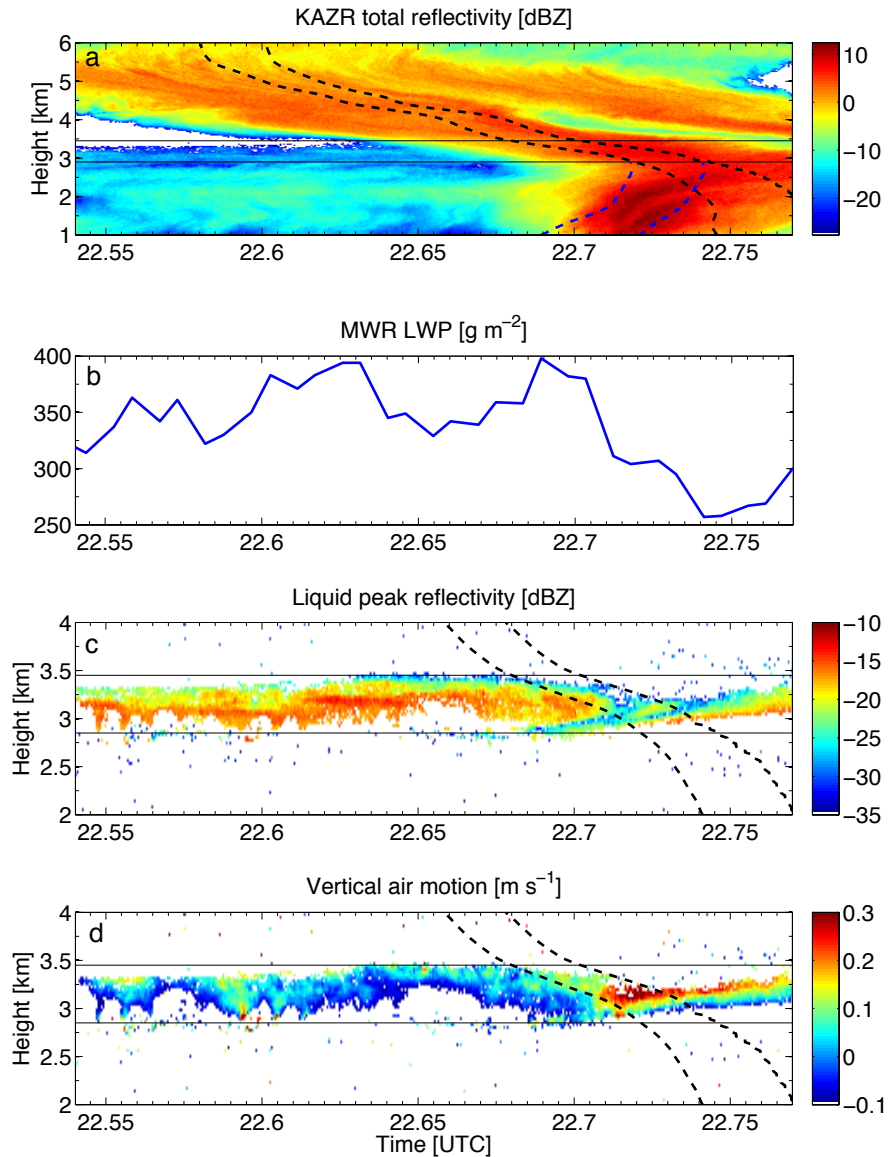
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4 Figure 4: Examples of KAZR Doppler spectra at different times and heights. Notice the
5 different x- and y-scales. Maximum and mean noise floor determined according to Hildebrand
6 and Sekhon (1974) are indicated by grey and black horizontal lines, respectively. Primary
7 peak is labeled as 1st, secondary peak as 2nd. Mean Doppler velocity of primary peak is
8 shown by vertical grey line. The plots are created with the Doppler Spectrum Visualizer, a
9 visualization toolkit which is publicly available at
10 http://www.gim.bnl.gov/armclouds/specvis_java_toolkit/

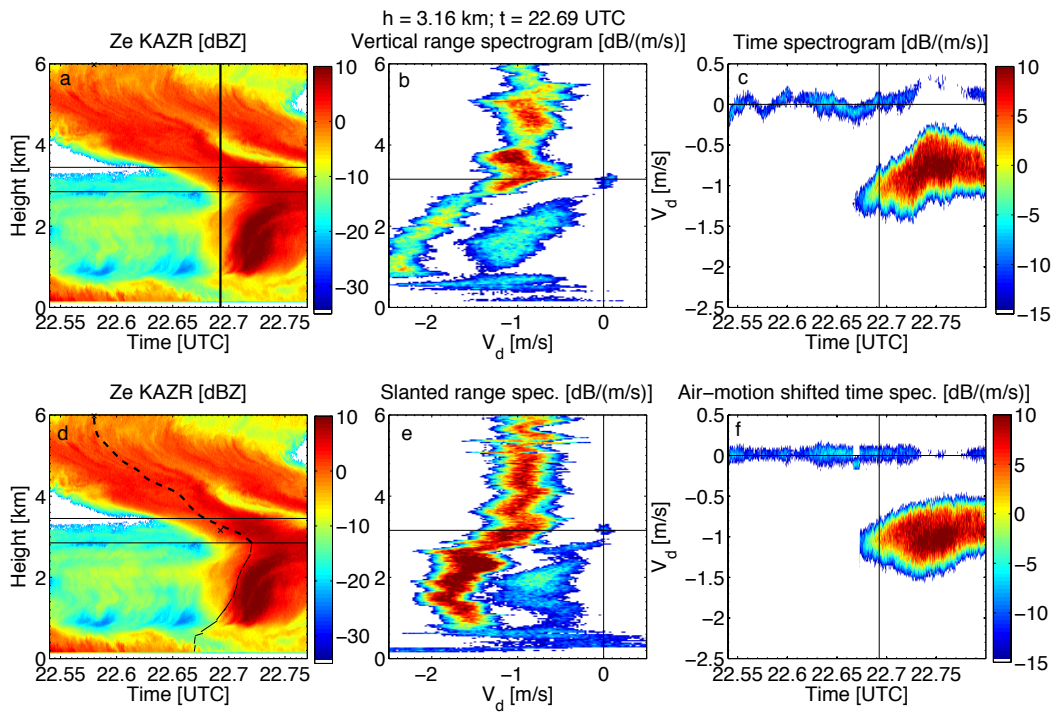




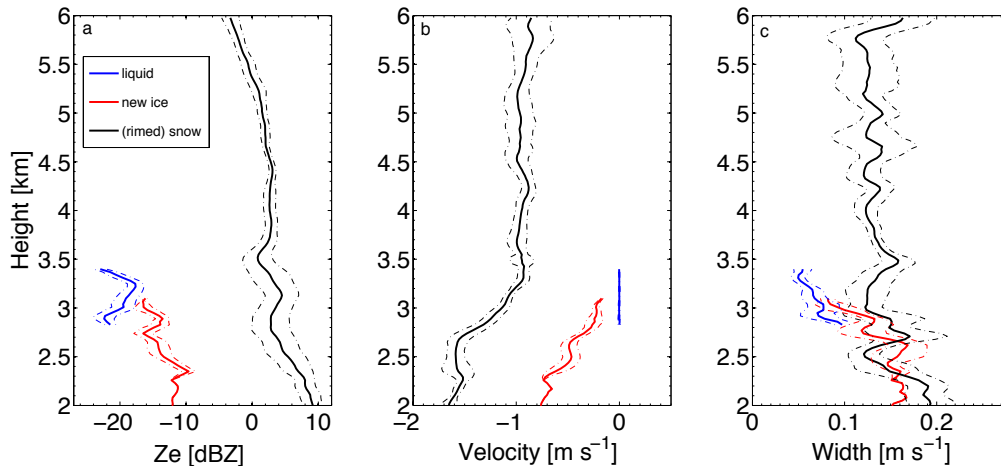
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2 Figure 5: Zoomed view (2-4 km, 22.54-22.77 UTC) of the snow front on February 21, 2014.
 3 Panel a) shows KAZR total reflectivity, b) microwave radiometer (MWR) liquid water path
 4 (LWP), c) reflectivity of the liquid peak, d) mean Doppler velocity of the liquid peak used as
 5 vertical air motion tracer. The dotted lines depict individual fall streaks starting at different
 6 generating levels z_{gen} ($z_{\text{gen}} = 6$ km (black), $z_{\text{gen}} = 2.9$ km (blue)). All subsequent averaged
 7 profiles refer to the area between the two black fall streaks.

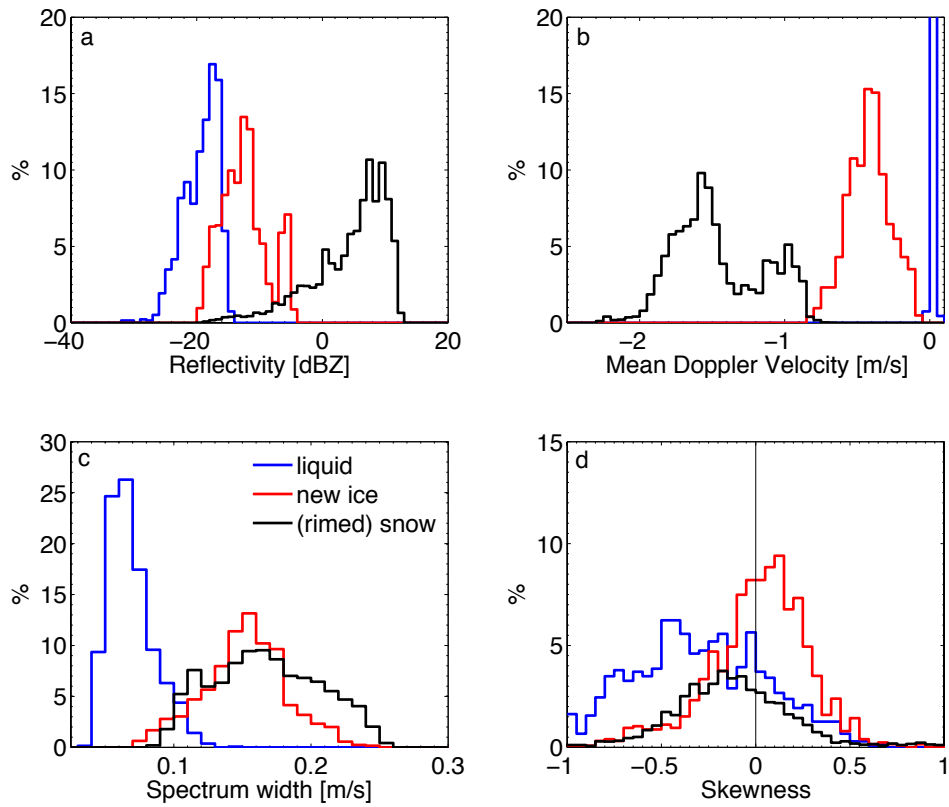
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 2 Figure 6: Panel a) and d) show KAZR reflectivity (Ze) field. There, the location of the SLW
 3 layer is indicated by thin horizontal black lines. Upper middle and upper right panel show
 4 range- and time spectrogram of vertical profile which is indicated by black vertical line in a),
 5 lower middle and lower right panel along slanted fall streak shown in d) as dashed line. b)
 6 vertical range spectrogram along vertical black line in upper left panel, c) time spectrogram at
 7 3.16 km, e) slanted range spectrogram along dashed line in d). f) air-motion corrected time
 8 spectrogram at 3.16 km.
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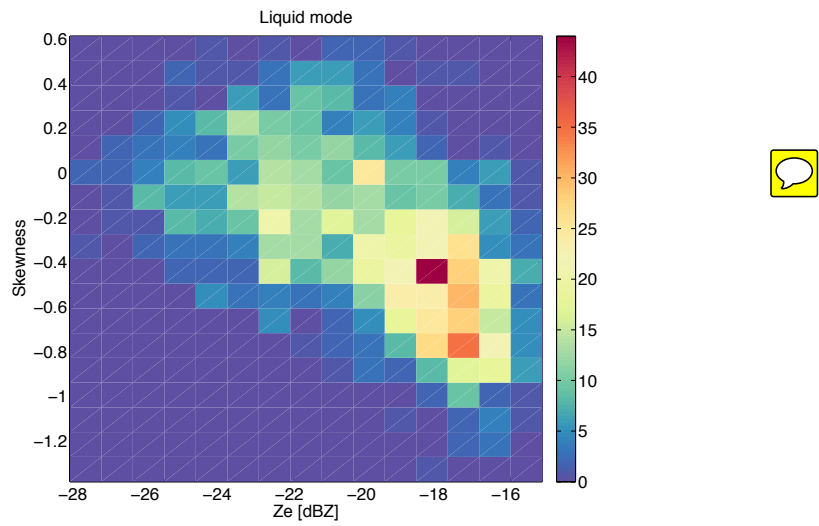
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 2 Figure 7: Slanted profile moments of SLW (blue), ice generated in the SLW layer (red), and
 3 frontal snow (black). Means and standard deviations of a) reflectivity, b) mean Doppler
 4 velocity, and c) mode width are shown for the slanted paths encompassed by the two dashed
 5 lines in Figure 5. Within the SLW water layer, velocities are corrected for vertical air motion.
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2 Figure 8: Probability density functions (PDF) of noise-separated peaks of SLW (blue), ice
 3 generated in the SLW layer (red), and frontal snow (black) for the slanted paths encompassed
 4 by the two dashed lines in Figure 4. Only data from surface to below top of the SLW layer
 5 (3.4 km) is considered. Please note that the PDF of SLW V_d is cut at 20%.

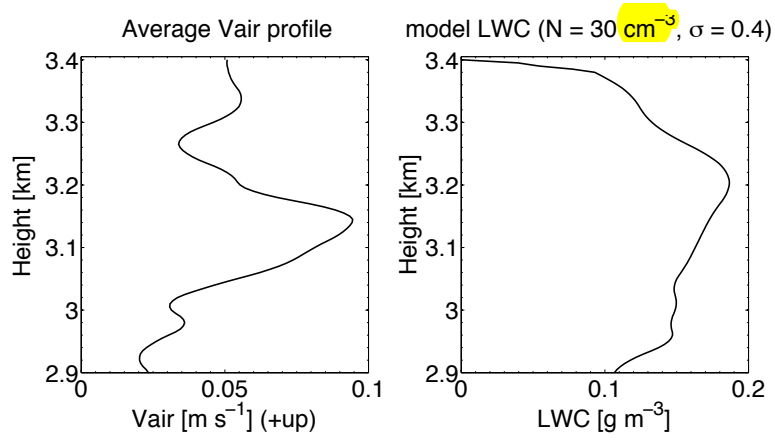
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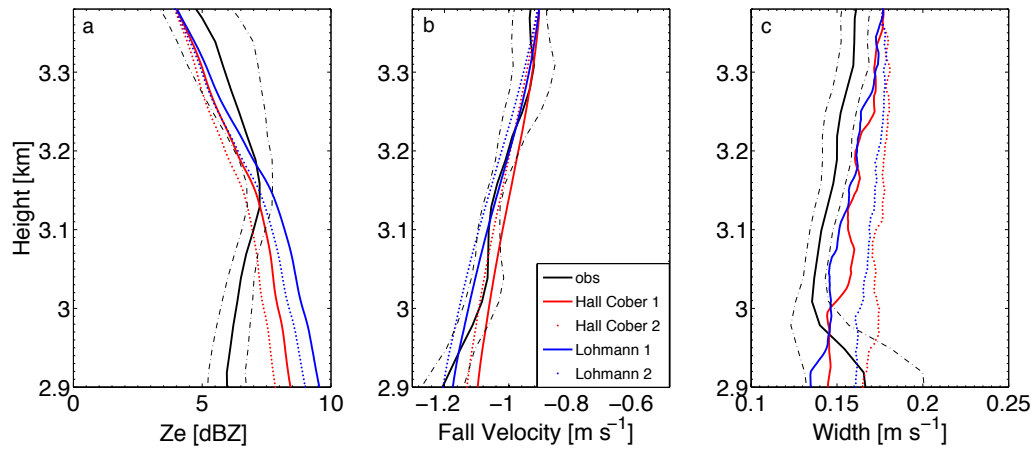
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2 Figure 9: Joint PDF of frequency of occurrence of skewness and reflectivity of supercooled
 3 liquid mode encompassed by the two black fall streaks in Figure 5.

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- 2 Figure 10: Mean vertical air motion profile (Vair, left) derived from liquid mode mean
- 3 Doppler velocity and liquid water content (LWC, right) derived from liquid peak reflectivity
- 4 between the two black fall streaks in Figure 5, respectively.
- 5



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 2 Figure 11: Comparison of observed and modeled moments of the frontal snow mode in the
 3 SLW layer. Black solid and dashed lines refer to the mean and standard deviation of the
 4 slanted paths encompassed by the two black fall streaks in Figure 5. Model results based on
 5 riming efficiencies by Hall & Cober as well as Lohmann are shown in red and blue,
 6 respectively (for details see text). Model results assuming area ratio increase with mass and
 7 rimed fraction (solid red and blue line) as well as with mass only (dotted red and blue) are
 8 shown.