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Measuring Snow with Weather Radar

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1. Introduction

People of warm climate tend to think of snow as something rather rare and exotic. However, most weather radars operating at mid-latitudes measure snow every day, at some altitude. Even at relatively low elevation angles the edges of a PPI image are often measured above the freezing level. Fig. 1 shows radar measurements from a night when surface minimum temperature was +13 °C – still, the majority of radar measurement volume was filled with snow.

In many meteorological classifications hydrometeors are divided in two or three classes (rain, wet snow and dry snow, see Fig. 1.). On the other hand, we have been told that there are not two identical snowflakes. Between these extremes are the snowflake type classifications such as those by Ukichiro Nakaya (Nakaya, 1954). From his work we can learn how, based on temperature and humidity, snow crystals can take the shape of needles, columns, plates, stars, rosettes and dendrites, only to name a few. They can also join each other in a process called aggregation, and they can be covered in icing in a process called riming.
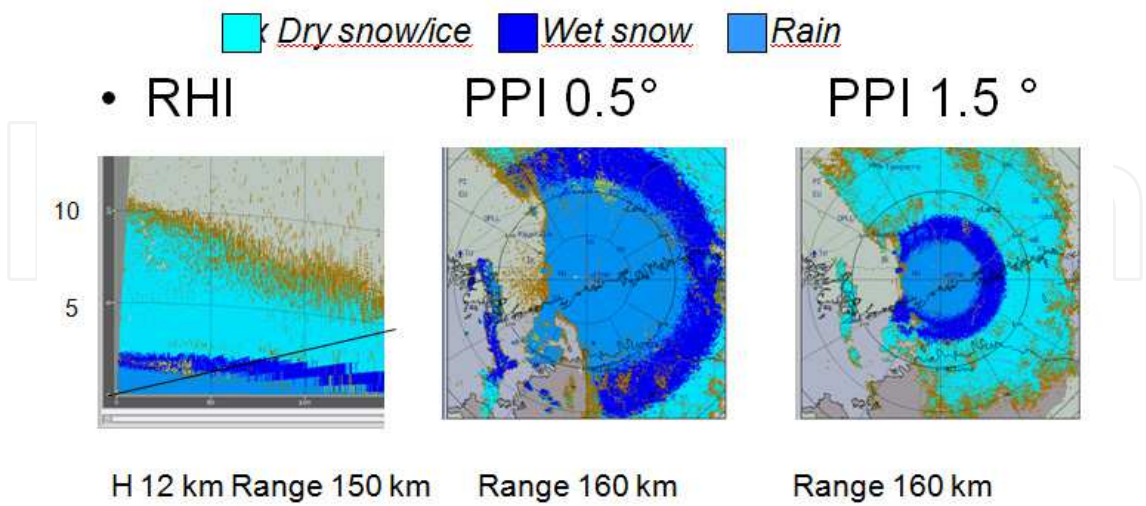

Fig. 1. Hydrometeor classification 30 August 2009 00:45 UTC in Vantaa, Finland. RHI to 150 km in range, 12 km in height (left). PPI to 160 km in range, 0.5 degrees in elevation (middle) and 1.5 degrees in elevation (right). Cyan for dry snow, dark blue for melting snow, light blue for rain.

Measuring snowfall with short wavelengths can bring us to the edge of assumption of the radar equation for Rayleigh scattering: are the particles much smaller than the radar wavelength ?

In this chapter, snow will be discussed from viewpoint of a radar meteorologist. Many topics are relevant for operational weather service, others more for the researcher. Increasing use of polarimetric radars is bringing new perspectives to measuring snow with radars.

2. Vertical structure of snowfall

With the usual measuring geometry of a scanning weather radar we have to take into account the vertical structure of precipitation. In warm weather, we measure rain near ground, wet snow above it and dry snow on the top, as can be seen in Fig. 1. A typical reflectivity structure is related to the temperature structure so that we have a maximum just below 0 °C isotherm. Above it, in the snowfall area, reflectivity decreases with an even gradient of approximately 7.5 dBZ/km. This decrease is related to four factors:

- at higher altitudes, it is colder and snow crystals are typically smaller in diameter
- at higher altitudes, the absolute humidity is smaller so the mass of snow per cubic kilometer of cloud is smaller there
- crystals fall down while they grow, so older crystals which have had time to grow large are more likely to be located at lower altitudes
- near the cloud top there may be effects of partial beam overshooting

In the precipitation system of a warm front, the two first factors create also horizontal gradients: the leading edge is in colder and drier air.

When the snowflakes melt, the surface gets wet first while the inner parts are still of dry snow. The partially-melted, wet snowflakes have approximately the size and fallspeed of snowflakes, but the dielectric properties of water surfaces. Hence the radar reflectivity peaks in the melting layer, a phenomenon also known as the bright band. In the hands of an inexperienced user of radar data, this could lead to an overestimation of precipitation intensity. In a modern weather radar service, the overestimation is corrected using knowledge of the vertical profile of reflectivity (Koistinen et al., 2003). Recently, Giangrande et al. (2005) and Boodoo et al. (2010) have shown, that the parameters of dual-polarization radars can be used effectively to follow the temporal and spatial variation of the melting layer height and thickness. This is especially important in cold and temperate climates, where much of precipitation is associated with fronts, because in frontal situations the temperature gradients are sharp.

In Fig. 2 we see RHI and PPI images in a snowstorm in Finland 2 February 2010. Cloud tops are observed between 6 and 8 km, and reflectivity is growing downwards from there. No bright band is observed, as there is no melting. Temperatures in cloud tops are near -35..-40 °C, at ground -5..-7 °C (based on Tallinn and Jokioinen 00 UTC soundings). The effect of vertical gradient is obvious in the RHI image, but the gradient in PPI is related to two factors: the vertical gradient and the horizontal variation of intensity.
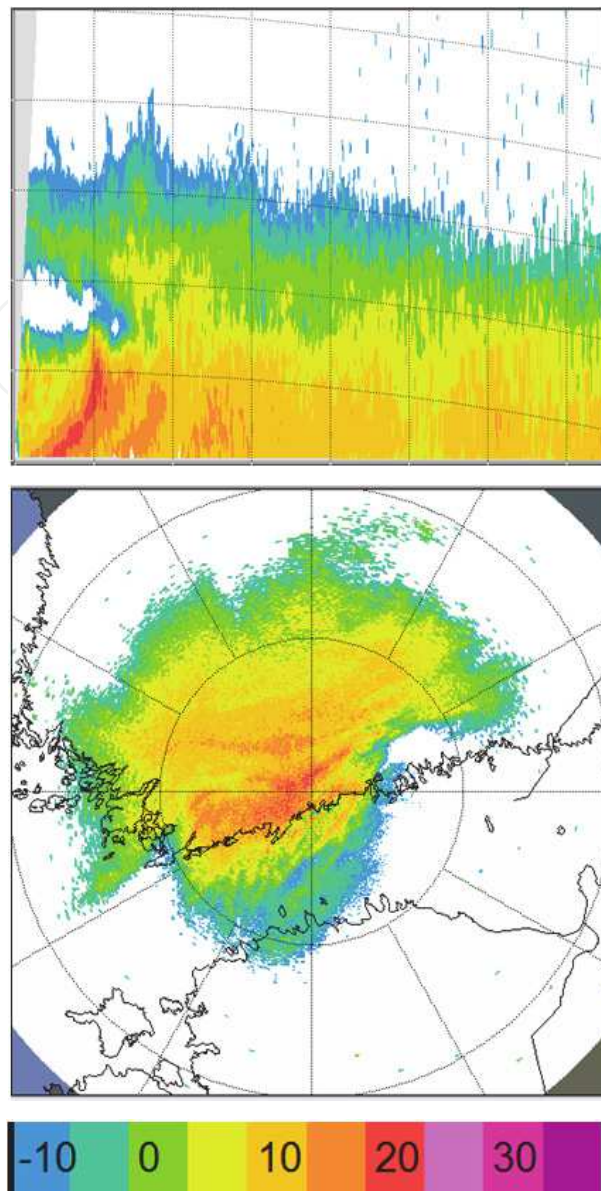
Fig. 2. Radar measurements in a snowstorm in Finland 2 Feb 2010. RHI north of radar to range of 150 km, horizontal lines at every 2 km (upper panel) and PPI with elevation of 1.5 degrees, rings at each 100 km (lower panel). Reflectivity values are between -15 and +25 dBZ.

3. Fallspeeds of solid hydrometeors

Terminal fallspeeds of hydrometeors are influenced by their type and size, and hence Doppler spectra measured with vertically pointing radar can be used for hydrometeor classification. Barthazy and Schefold (2006) showed that the fall velocity of snowflakes consisting of needles or plates is strongly dependent on the riming degree. The average fall velocity of any type of snowflakes of diameter of 1 mm or larger is typically between 1 and 2 m s^{-1}. In cases when hydrometeor types change, or two types of hydrometeors coexist in same measurement volume, Doppler spectra can provide valuable information of cloud physical processes such as riming and aggregation. In addition of academical interest they may provide value in aviation weather services.

4. Clutter cancellation and clear air echoes

One of the main reasons why operational weather services started to use Doppler radars was the use of Doppler signal for clutter cancellation of reflectivity fields. The principle is simple: precipitation has velocity (at least fallspeed and turbulence), while ground clutter does not, and Doppler radar can measure the velocity (or absence thereof). However, when the precipitation is in form of snow, there are some complicating details, and we have to study the filtering process in depth.

A Doppler radar does not measure the true speed of the particles, just the component parallel to the radar beam. When the real wind is nearly perpendicular to the beam (e.g. for northerly winds in east and west of radar), this component is near zero. We can not set the threshold to censor only the bins with exactly zero speed, because even clutter targets have apparent speeds due to different viewing angle of the rotating antenna, trees and masts waving in wind and trucks and lorries in an urban environment. On the other hand, if the threshold of censorship is too high, Doppler filter removes data with near-zero Doppler velocities in areas where wind is perpendicular to beam. Because there is wind shear and measurement is made at different heights at different distances, the direction of missing data is undulating, and hence the gap in reflectivity PPIs is sometimes called *Doppler snake.*

In Fig. 3, it is relatively easy to see the underestimation of reflectivity on a line coiling from west-southwest through the radar and to the opposite side. In this case, there is also a "secondary snake" south of the radar, where the near-zero-velocities are related to folding.
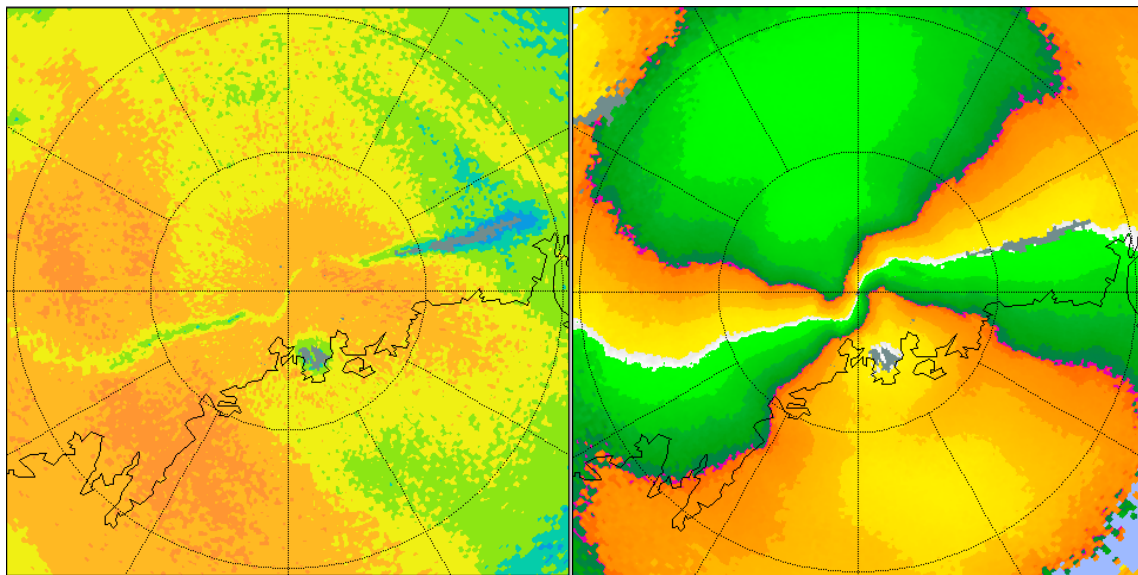

Fig. 3. A Doppler snake case 17 March 2005 03:00 UTC. PPIs of reflectivity (left) and Doppler velocity (right). Wind is from south-southeast, warm colours indicate echoes moving away from the radar and cold colours towards the radar . The velocity field is folded, unambiguous speed 7.6 m/s. A band of weaker reflectivities cutting through the image from southwest to east, at same location as zero Doppler speeds (white). Reflectivity colour scale as in Fig. 2.

It is tempting to try to get rid of the Doppler snake by using a less aggressive Doppler filter. However, amount of residual clutter may increase. Finding the compromise between too aggressive and too weak filter is threading on a fine line, and the selection should be tested in different wind and temperature conditions. Temperature inversions (typical for cloudfree winter days or nights) affect amount of clutter by causing anomalous propagation, which then leads to the increase of ground or sea clutter. Wind affects sea clutter but may also cause blowing snow. Especially the blowing snow falling from trees in hilly areas may give false alarms of ground clutter – it is real snow flying in real wind and hence immune to most clutter cancellation techniques, even though it is not precipitation falling from clouds.

Case of sea clutter is especially annoying. In summer we have nocturnal inversions and anomalous propagation mainly when it is not raining. In winter it is very likely to have on continent cold weather and inversion, and simultaneously rigorous lake effect snow (see section 6.2) over the water areas. This is possible, because the propagation is affected by temperature near the radar, not at the measurement location. Sea clutter is immune to Doppler filtering, but dual polarization measurements can reveal it, as is seen in Fig. 4.
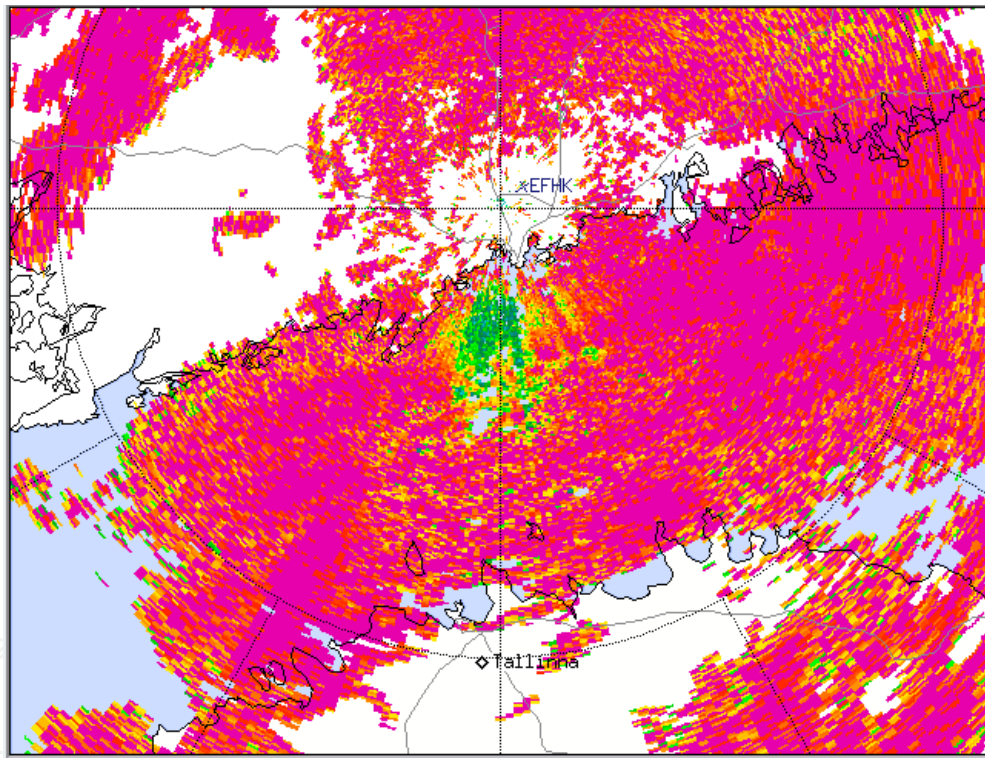

Fig. 4. Sea clutter and lake effect snow seen with polarimetric parameter RhoHV. RhoHV over 0.98 in snow, less than 0.8 in sea clutter. The range ring indicates 100 km from the radar.

5. Dual polarization

While one of the most popular applications of dual-polarisation technology enables one to distinguish the types of the hydrometeors measured (Straka et al., 2000), dual polarization can be used for much more. On the other hand, some applications developed for rain, such as KDP-based algorithms for quantitative precipitation estimates do not work in snow.

When implementing published algorithms to new environments, it is wise to compare the hardware used in the original development work to the platform where it will be implemented. Research community has used S-band radars a lot, while the operational weather services in cold climates use mainly C-band. There may be also difference between simultaneous and alternating transmission of the dual polarization channels.

5.1 Hydrometeor classification

Polarimetric properties of wet snow and single snow crystals are very different from the ones in rain. Hence, these two snow categories are easily distinguishable from rain. However, discrimination between stratiform rain and dry snowflakes (aggregates) is challenging, as relatively low Z and ZDR, and high RhoHV are typical for both (Ryzhkov & Zrnic, 1998). A typical solution in this case is to use the information of height of melting layer, either from sounding, NWP model or radar measurements.

Decision boundaries for dense snow, dry snow, wet snow, rain, dry graupel, wet graupel rain and hail co-existing and hail alone using Z, ZDR, RhoHV, LDR and KDP have been published by e.g. Straka and Zrnic (1993). In many of the parameters, the selected classes overlap, which has encouraged researchers to try fuzzy logic (Bringi & Chandrasekar, 2001). In Fig. 5 we see RHI scans in the same situation as in Fig. 1. The 0 °C isotherm is at 2 km, and a layer of melting snow can be seen below it.
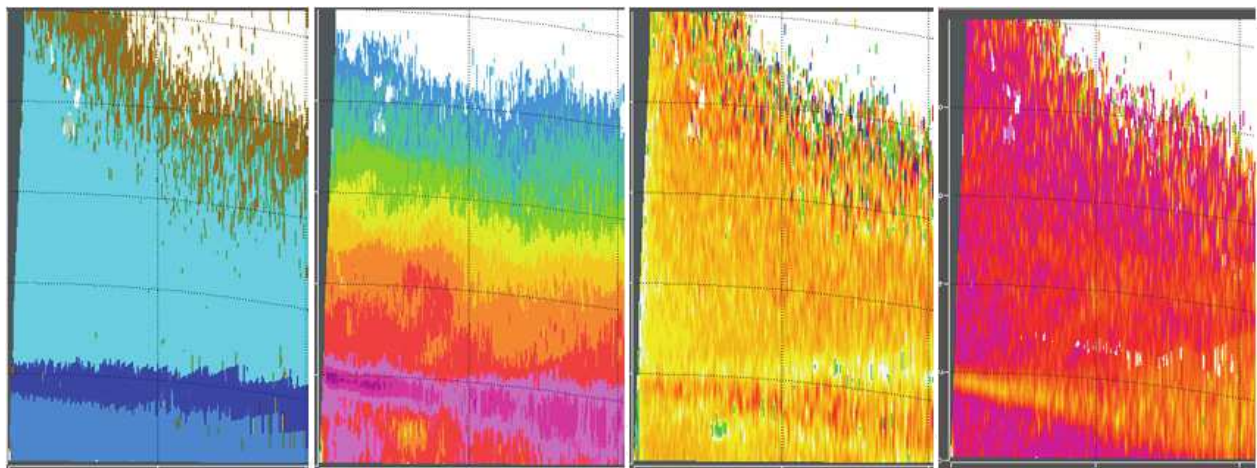

Fig. 5. RHI scans north of Vantaa radar 30 August 2009 00:45 UTC. . Range 100 km, height 10 km, parameters from left to right: Hydrometeor classification, reflectivity Z, differential reflectivity ZDR and copolar correlation factor RhoHV.

5.2 Snow types

In the JPOLE classifier for cold season Ryzhkov et al. (2005) distinguished between Dry aggregated snow: DS and Wet snow: WS, and single crystals. In some cases, this distinction can be made using reflectivity and differential reflectivity alone: wet snow has typically larger reflectivities, and individual crystals larger differential reflectivities. However, the variability of both parameters for both classes is large, and the definitions tend to overlap.

Developing a universal method for snow type classification is even more challenging than finding a representative ZR relation: reliable surface observations of snowflake type are rare, and if they are performed at longer distances from the radar, the snowflakes can change between the radar measurement and the surface observation. Correction for vertical profile of reflectivity is a standard procedure, but correction for vertical profile of snow type is still strongly hypothetical.

6. Characteristic properties of typical snowfall situations

In general, precipitation events can be split to orographic, frontal and convective precipitation. Especially snowfall is often related to warm fronts and lake effect induced convection.

6.1 Warm fronts

Much of snowfall is related to frontal systems of extratropical systems. In their analysis and forecasting, the value of weather radar data lies primarily in the mesoscale structure: detection of the mesoscale bands of heavy snowfall is needed for accurate short term forecasting. The banded structure leads to rapid changes in visibility, and areal differences of accumulated snowfall. Their dynamical structure is complicated, related to negative equivalent potential vorticity (EPV) mainly associated with conditional symmetric instability (CSI), and not always perfectly forecasted by numerical weather prediction models. Hence, identification and extrapolation of movement of these bands using a radar can improve short-range forecasts of extreme events significantly (Nicosia and Grumm, 1999).

Snowfall from warm fronts is also a challenge for a radar meteorologist: forgetting the three-dimensional structure of the frontal system can lead to embarrassing misinterpretation.

In satellite images, we can see the leading edge of frontal system ("warm front shield") and educated meteorologists already know, that arrival of this edge does not mean onset of precipitation. In Fig. 6 the shield at 2 km extends 70 km ahead the surface precipitation. I sincerely hope that everyone using different radar products remembers this, too: the leading edge in products like TOPS, MAX, VIL or even medium-level CAPPI does not indicate the precipitation on ground level. See Fig. 7.

The sloping edge of precipitation area can also be seen in PPIs. In upper panels of Fig. 8, the gap in the centre of the image indicates area where radar beam was below the warm front shield. The gap gets smaller when the surface front approaches the radar. Also, the gap is not a circle but an oval, also indicating the slope of the cloud base.

Warm fronts are ideal for producing Doppler wind profiles (VAD and VVP), because the wind field is usually uniform. In lower right panel of Fig. 8 we have a time series of VVP wind and reflectivity. In this case, the wind shear related to the warm advection is not very strong, and it is hard to distinguish it from wind shear related to friction in the boundary layer. Sharper turning of wind is of interest to aviation weather service. We can also see the shield of overhanging precipitation with approaching front (05-06 UTC) from "+"-signs indicating missing wind barbs. Warm advection can also be seen indirectly: part of the increase of reflectivity at low altitudes around 08 UTC is probably related to temperatures rising to near zero, and snowflakes growing larger.
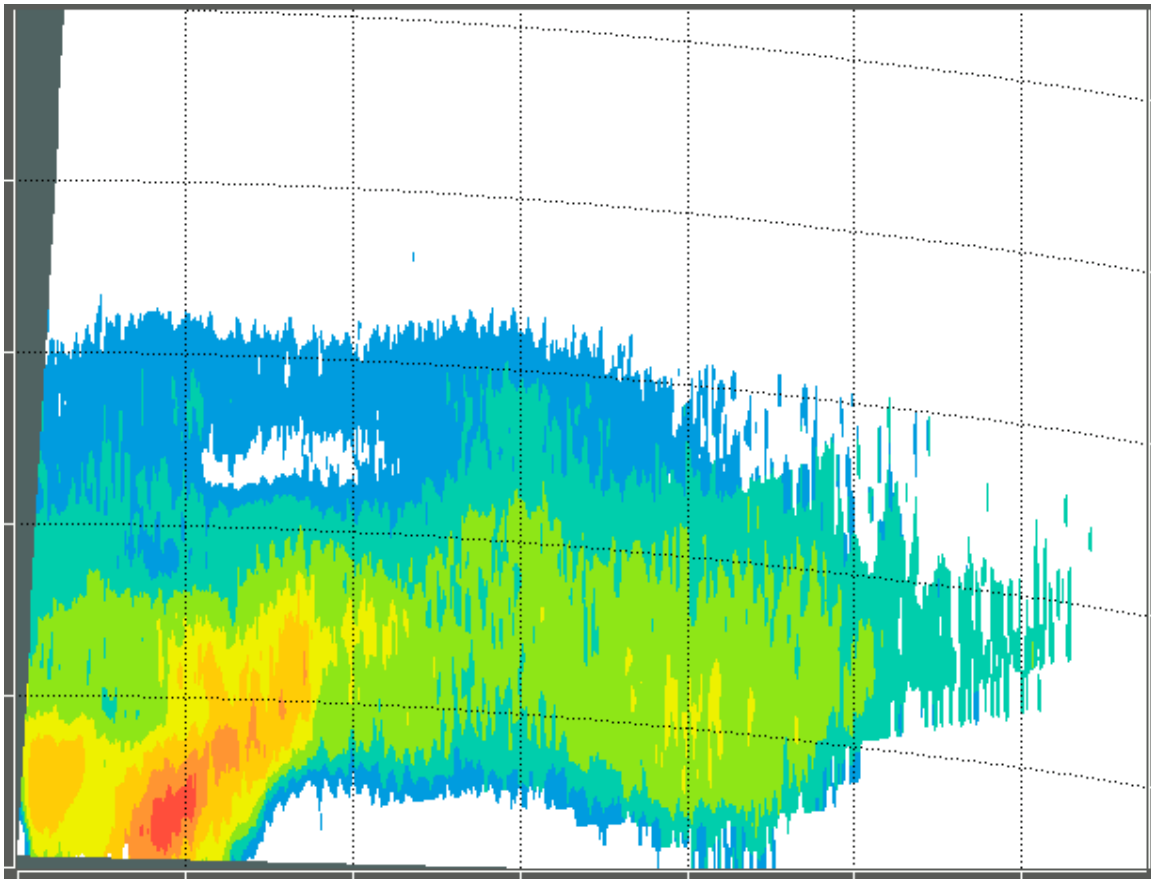
Fig. 6. RHI north of Anjalankoski radar 17 December 2011 12 UTC, warm front approaching from south. Vertical lines at 20 km, horizontal lines at 2 km intervals. Colour scale from -10 dBZ (blue) to 20 dBZ (red) as in Fig. 2.
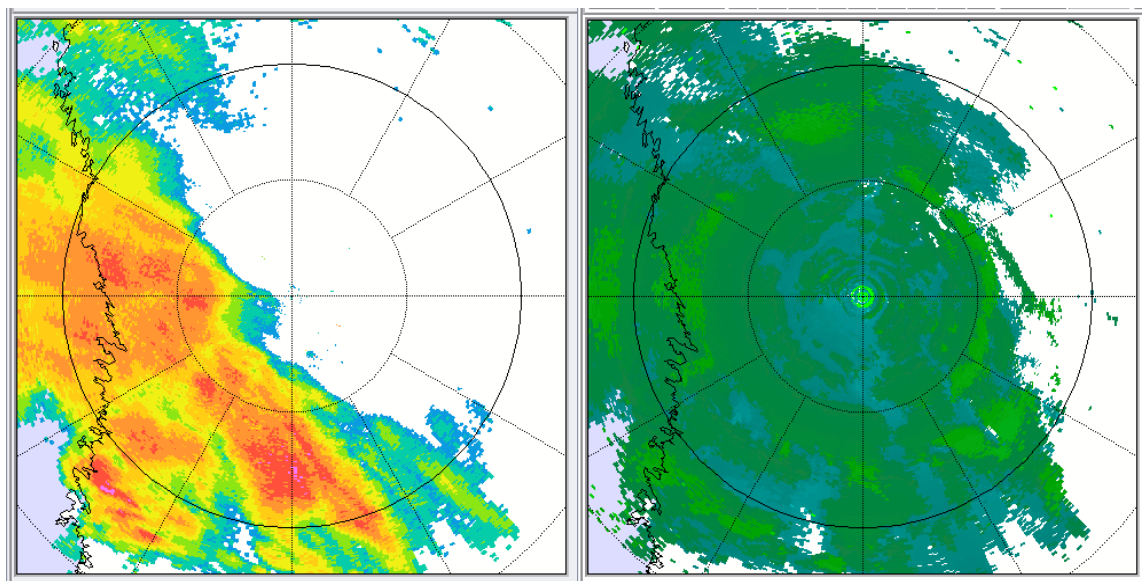

Fig. 7. CAPPI at 500 m height on left, TOPS with threshold -10 dBZ on right. 14 January 2008 05:30 UTC, precipitation on the surface had not yet reached the radar location. Range rings at 50 and 100 km from radar, reflectivity scale as in Fig. 2., green shades for tops in steps of 2 km.
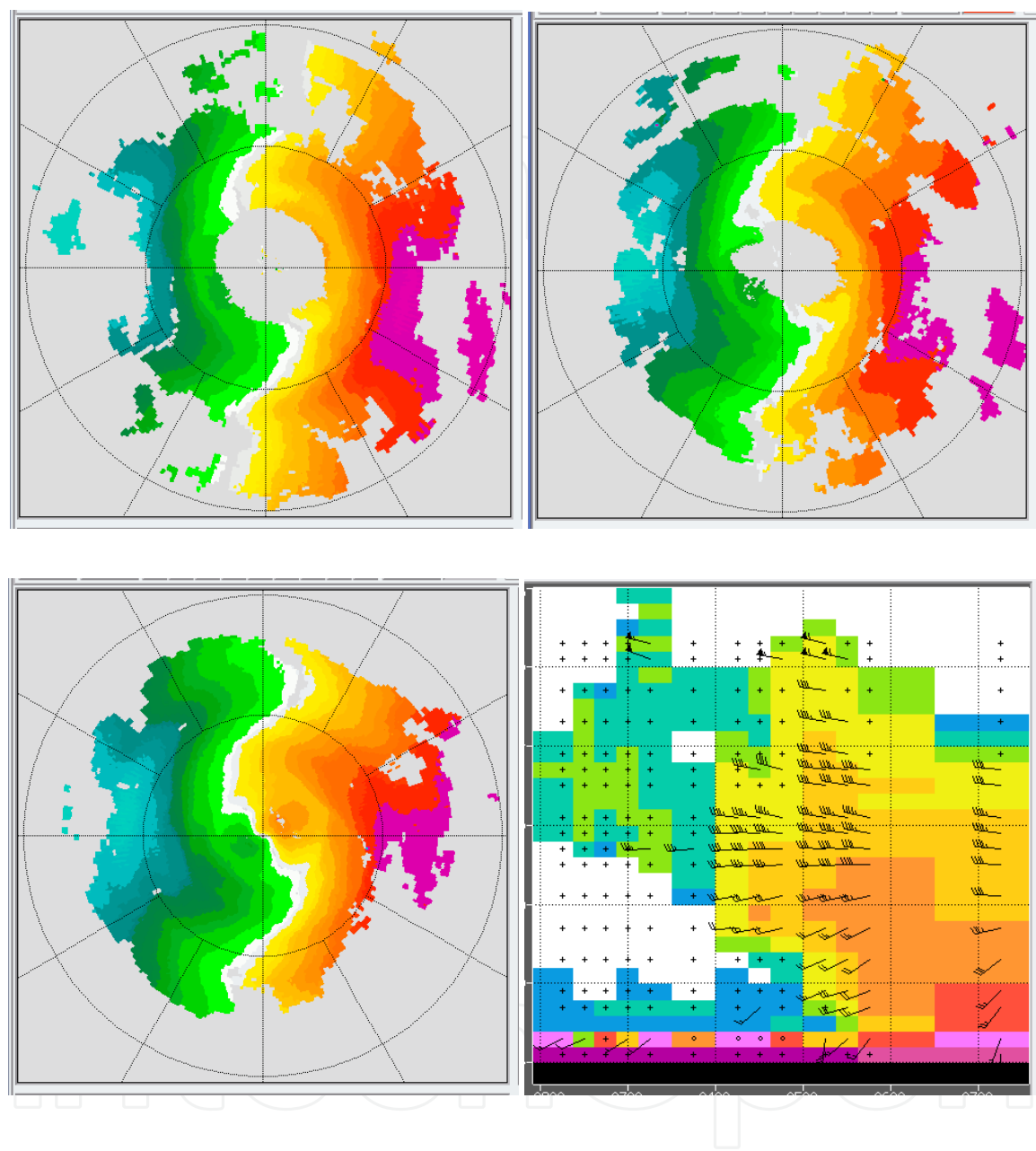
Fig. 8. PPI images of Doppler velocity 30 minutes apart (5:00, 5:30 and 6:00) and time series of wind and reflectivity profiles 02:00 to 07.00 UTC 14 January 2008. Green colours towards radar, reds and yellows away. Reflectivity from -20 dBZ (blue) to +20 dBZ(purple)

In Fig. 9 we see another scale of shear. There is a small wind maximum below 1 km. In this case it lasted for less than 45 minutes. This is an indicator of low level jet related to conveyor belt in the frontal structure. In the real atmosphere, the wind shear related to that is probably larger, as the VVP is averaged over a cylinder of 30 km.
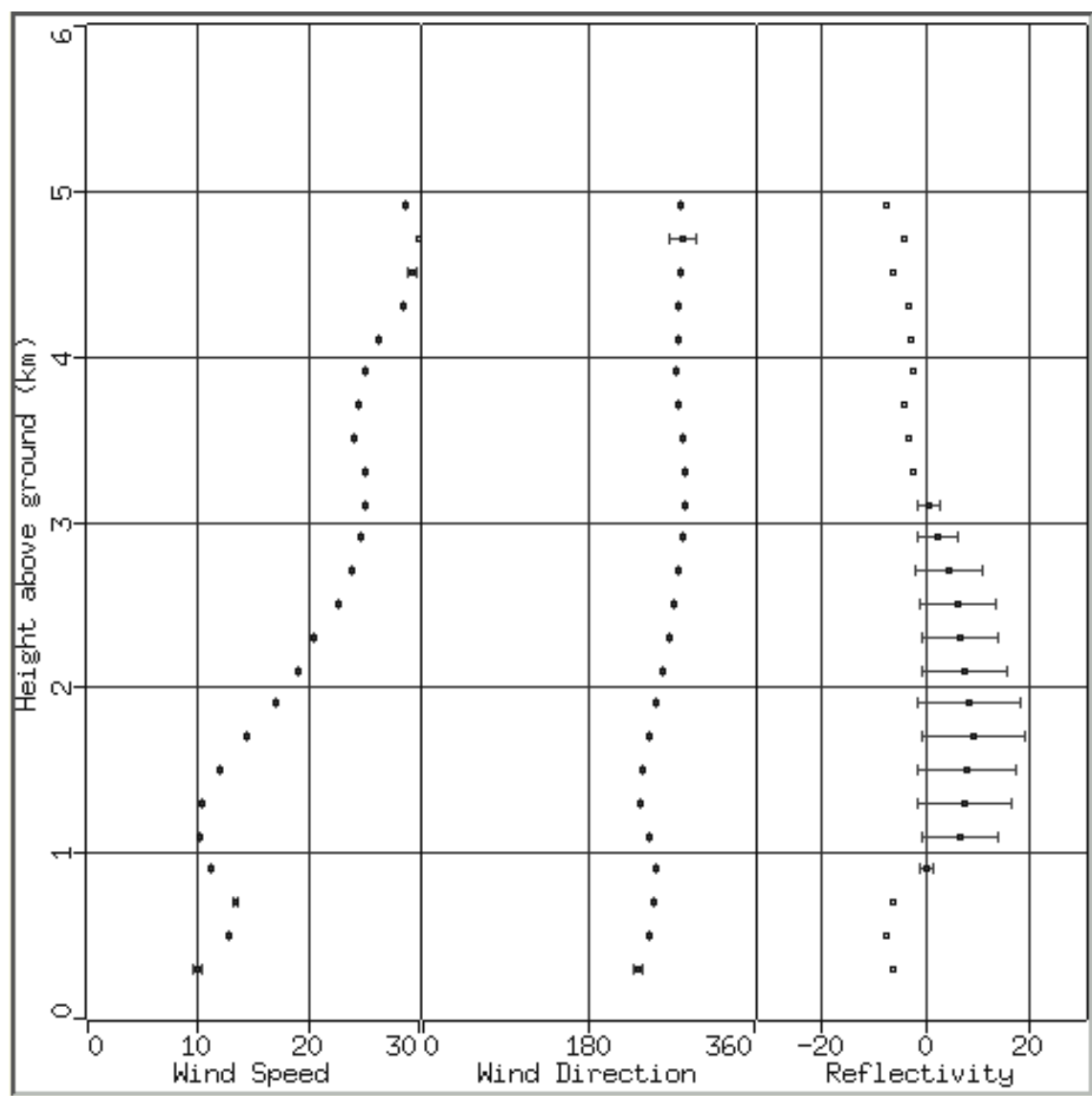
Fig. 9. VVP sounding. Windspeed on left, wind direction in middle and reflectivity on right.

6.2 Lake effect snow

Lake effect snow is a phenomenon observed regularly around open water surfaces in cold weather. The name originated from weather phenomena around the Great Lakes of North America, but it is also observed around bays and straits of sea and great rivers. Lake-effect snowstorms get their energy from the temperature difference between the relatively warm open water and very cold, continental air blowing over the water. These provide the most spectacular outbreaks of boundary layer convection in winter (Markowski & Richardson, 2010).

Markowski and Richardson (2010) mention that the convective clouds associated with lake-effect precipitation can be several kilometres deep. However, even shallower lake-effect clouds can produce significant amounts of snowfall (see Fig. 10), and these shallow yet intense clouds present challenges to the design of a radar network in coastal areas in a cold climate.

The organization of convection in a lake-effect snowstorm depends on the ratio of the wind speed to the maximum fetch distance. When the wind is strong, offshore convection is rapidly organized into horizontal convective rolls. When the wind is weaker, it is more likely that bands parallel to the shoreline (and perpendicular to the mean wind) are formed in the land-breeze convergence zone. With very weak winds, convection can be organized into vortices that stay over the sea and have the structure of a miniature hurricane (Laird et al., 2003). These vortices provide another example of mesoscale weather systems which can only be observed with remote sensing instruments.
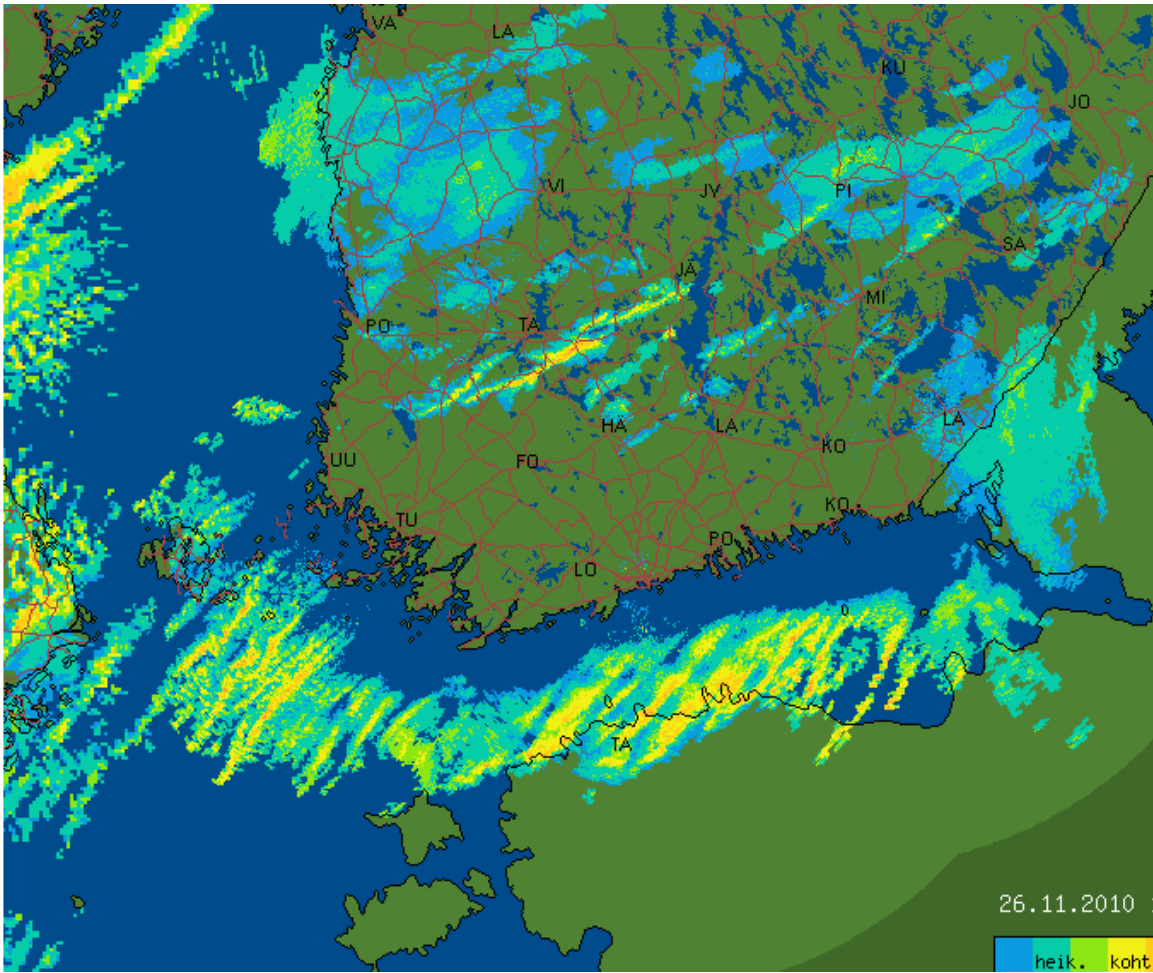

Fig. 10. Lake effect snow 26.11. 2010 10 UTC. Pseudo-CAPPI reflectivity composite of 5 Finnish and 2 Swedish radars at nominal height of 500 m.

For radar-based nowcasting applications, lake effect snow is a challenge firstly, because the snow storms do not move, thus motion vectors can be misleading, and secondly because their shallow and intense nature can cause beam overshooting problems..

7. Operational applications

Snowflakes are beautiful and interesting, but most people who buy radars want also to do something useful with the data. The most frequently asked questions are when, where and how much. These have been solved with more and less advanced accuracies over the years. Advanced questions to be still researched are related to properties of snow, most of all its density.

7.1 Accumulated snowfall, Z/S

When we talk of precipitation, we often think about rainfall, and ignore snow. There are two main approaches to accumulated snow: 1) snow water equivalent (SWE): how large is the mass of snow per unit area and 2) the (increase of) thickness of snow layer (TSL). SWE is important for hydrological applications, and it is also usually the parameters measured at surface station rain gauges (in manual gauges, the snow is melted and volume of resulting water is measured). SWE has also applications in estimates of snowload of buildings and tree crowns. Thickness of snow layer has applications in road maintenance, biology and recreational activities, and it is a tricky thing to estimate for everyone, not just radar meteorologists. A bulk equation TSL=10*SWE (both TSL and SWE expressed in mm), is often used, even though we all know that the density of snow on ground varies a lot. Matters are further complicated if we try to accumulate TSL over a longer period, because snow on ground is changing shape, density and even location (by blowing snow).

The radar equation (see Zrnic, this volume, for details) includes the parameter $|K|$, dielectricity of scattering particles, which has different values for ice and water. In operational signal processing, this is assumed to be always the water-value, so we should call the measured parameter Ze (where the "e" stands for "equivalent"). The error caused by assuming the same dielectricity is compensated in using a different ZS relation for snow, and including the effect of dielectiricity there.

The density of snow crystals and aggregates varies as a function of structure from 50 to 900 kg m^3, with higher values expected for solid ice structures and wetted particles. The size distributions of ice crystals and snow aggregates can be represented by exponential and gamma functions, and the total number of concentrations is on the order of 1–10^4 m^{-3} for aggregates, 10–10^9 m^3 for individual crystals at colder temperatures (T < -20 °C), and often as high as 10^4 m^{-3} at warmer temperatures (Pruppacher & Klett, 1996). The diameter of large crystals can be up to 1–5 mm, while the diameter of aggregates can grow to 20–50 mm, occasionally even more. The shapes of aggregates vary from approximately spherical to extremely oblate, and the approximate shapes of crystals can vary from extreme prolates and oblates to essentially spheres (Pruppacher & Klett, 1996). Most individual crystals tend to fall with their largest dimension horizontally oriented unless there are pronounced electric fields. Aggregates also can fall in a horizontally oriented manner or may tumble (Straka, 2005).

Radar estimates of ice water content of crystals and aggregates are greatly complicated by the multitude of crystal sizes and shapes, various crystal and aggregate densities, and dielectric constants, among others. Of all the snow types, the determination of the amount of wet aggregates is probably the most difficult (Straka, 2005).

For liquid precipitation, the classical ZR relation was published by Marshall and Palmer 1948. For snow, similar classical paper is probably that of Sekhon and Srivastava (1970). However, when dropsize distributions are known to vary a lot, snow particle distributions vary even more. Applying correction for vertical profile of reflectivity before the ZR relation is crucial, but does not eliminate all uncertainties.

7.2 Nowcasting snow

The simplest application of radar images for nowcasting is to display a time series as an animation, and visually follow its speed and direction of movement. Second level of complication is to estimate the future movement with some vector field, which can be derived from observed movement, NWP, or even Doppler velocity field (note this is not recommended but some people do it). Compared to summertime convective precipitation, snow has some advantages and some disadvantages in this respect. Because snow is seldom related to convection, it has less diurnal variation, and hence frontal snowstorms can sometimes be tracked and extrapolated for several hours with fairly good accuracy. On the other hand, snowstorms tend to be shallow, and hence the geometrical factors can cause error in speed estimates, and even causes of total miss (snowstorms hiding under the lowest radar measurement).

Because snowflakes fall slowly, they can advect remarkable distances after the radar measurement. If we measure at height of 800 m, and the snowflakes fall 1 m/s, they reach the ground 800 seconds later, and if wind blows 10 m/s, the location can be 8 km downwind from the radar measurement. From height of 1800 m, the flakes fall for half an hour, and from height of 4 km more than an hour, and for a distance in order of 40 to 60 km. This affects all studies comparing radar measurements to "ground truth", and it can be annoying for nowcasting, too. On the other hand, for an optimist it is a source of information: basically, we have already measured the snowflakes which will fall e.g. to the runway half an hour later. Lauri et al. (2012) have discussed the effect of advection in snowfall measurement.

7.3 Visibility in snow

Aviation meteorology uses abbreviation LVP (low visibility procedure) and we often read this as "fog and stratus"). However, even snowfall reduces visibility in significant amounts. Unlike in fog, the visibility in snowfall often fluctuates rapidly and significantly, and hence use of radar data to aid nowcasting would be beneficial.

Visibility is related to scattering of visible light, radar reflectivity is related to scattering of microwaves. In case of particles in typical sizes of snowflakes and snow crystals, these two behave differently: if the amount of snow in air stays same, but crystals join to larger aggregates, radar reflectivity grows (following the ND^6 equation) while the optical visibility

improves (scattering gets smaller). Hence, any reflectivity-to-visibility –equations depend heavily on particle size distribution.

Rasmussen and Cole (2002) have given to visibility the equation Vis = k / ND², where k is a constant related to snow type. Having the two equations available, it would be tempting to "just solve the k" and get a reflectivity – visibility equation. However, the "constant" k can get plethora of values depending on the crystal type, the degree of riming, the degree of aggregation, and the degree of wetness of the crystals. Rasmussen et al (1999) derived ratios of visibility and liquid equivalent snowfall rate for 27 crystal types and two aggregate types. In their study, typical variations in visibility for a given liquid equivalent snowfall rate ranged from a factor of 3 to a factor of 10, depending on the storm.

After this, the next attempt would be try to "calibrate" the factor k for each storm. However, Rasmussen et al. (1999) also noted that k has a wide degree of scatter also during a given storm. As we know from other studies, the type of snow crystals depends on temperature and humidity it has experienced during its growth time. Snowstorms are often related to weather situations (such as warm fronts) with strong gradients of temperature and humidity. Hence, changes of crystal type during a storm are natural.

Another factor making comparison of reflectivity and visibility is the illumination. In same snowfall intensity, visibility in night (how far can you see a light source) can be twice as good as in daylight.

8. Conclusion

The four main properties a radar meteorologist should remember about snow are:

- It falls from shallow clouds, so you can't see it from far.
- It falls slowly, so it may advect after we measure it.
- It has different scattering properties from rain, so your precipitation estimates may be inaccurate if ZR relations are applied for snowfall
- Snowflakes come in many sizes and shapes, and their scattering properties may vary, so more information may be acquired using dual polarization.

9. Acknowledgment

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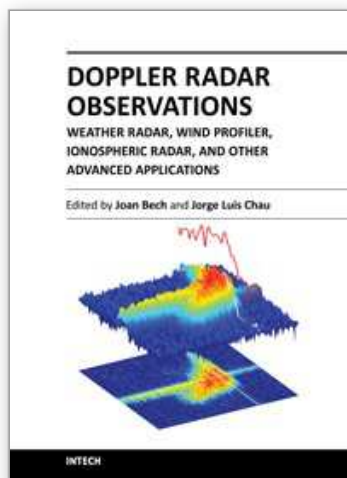
Doppler Radar Observations - Weather Radar, Wind Profiler, Ionospheric Radar, and Other Advanced Applications
Edited by Dr. Joan Bech

Doppler radar systems have been instrumental to improve our understanding and monitoring capabilities of phenomena taking place in the low, middle, and upper atmosphere. Weather radars, wind profilers, and incoherent and coherent scatter radars implementing Doppler techniques are now used routinely both in research and operational applications by scientists and practitioners. This book brings together a collection of eighteen essays by international leading authors devoted to different applications of ground based Doppler radars. Topics covered include, among others, severe weather surveillance, precipitation estimation and nowcasting, wind and turbulence retrievals, ionospheric radar and volcanological applications of Doppler radar. The book is ideally suited for graduate students looking for an introduction to the field or professionals intending to refresh or update their knowledge on Doppler radar applications.

How to reference

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