Spatial Continuity of Precipitation, Profiles of Radar Reflectivity and Precipitation Measurements in the Alps

A dissertation submitted to the Swiss Federal Institute of Technology (ETH) Zürich

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November, 2000
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Abstract

Studying Alpine precipitation is a difficult but stimulating task. The orography plays a key role in precipitation mechanisms, and thus in the distribution of precipitation in time and space. Weather radars provide unique four-dimensional observations of precipitation. But in the Alps, the mountains hide any precipitation behind them, and reflect the radar signal causing strong ground clutter. The result is a complex map of visibility, reflecting the lack of information in cluttered and hidden regions. Thus, interpreting Alpine radar data, in particular for quantitative use, requires optimum combination of the available observations with the knowledge of the visibility. In the three articles, reproduced in this thesis, we present

a) variograms of radar reflectivity describing the spatial continuity of Alpine precipitation,

b) a robust profile-correction scheme based on meso-beta profiles for improving precipitation measurements by radar in complex orography, and

c) quantitative estimates of attenuation caused by water on the radome.

The first article makes use of the high resolution of radar data in four dimensions. We use variograms of radar reflectivity to quantify the spatial continuity in different types of precipitation (stratiform versus convective, rain versus snow), and as a function of the position within the Alps (upstream, upslope, main divide). The variograms provide quantitative answers to practical questions related to the spatial continuity of precipitation, such as: To what accuracy can we estimate basin-average rates from point observations (gauges)? What difference do we have to expect between the precipitation rate at two points separated by a given distance, e.g., when comparing measurements of two instruments? How does the frequency and the type of convection depend on the orography? What is the uncertainty of single radar measurements?

Some results: For some selected events at a separation distance of 10km we found the expected difference of reflectivity to be 3.8dB(Z) in stratiform snowfall, 4.3dB(Z) in stratiform rainfall, 7.6dB(Z) in a meso-scale convective system (MCS), and 13dB(Z) during maximum convection of the MCS respectively. These values are for instantaneous measurements at about 20km range from the radar. They decrease when integrating reflectivity in time. Accordingly, the representativity of point observations for area-average rates depends on the type
of precipitation (time, and location within the Alps), as well as on the integration time. Take, for instance, a basin of 576km² and a period of 40min. Then, the fractional deviation between a point measurement and the basin-average ranges from 11% in stratiform rainfall, to 65% in convective rain.

The goal of the second article is to improve radar estimates of ground-level precipitation rates in the Alps by correcting for the vertical change of the radar echo (profile), caused by the growth and transformation of precipitation. To calculate the profile we integrate radar data on a meso-beta scale (a few hours × 140km). In the Alps, when ignoring the influence of the profile, radar measurements often underestimate ground-level precipitation by several dB (i.e. easily by a factor of 2). Occasional overestimations resulting from echo enhancement in the melting layer (bright-band) are of the same order of magnitude. The improvement of profile-corrections in terms of ground-level rates is assessed by comparing gauge measurements with four types of radar estimates: vertical maximum, lowest visible echo, estimates corrected by using the event profile, and estimates corrected with meso-beta profiles. Data for verification is from 247 hours of summer and winter precipitation. The radar-gauge comparisons show significant improvement obtainable with profile-corrections. The main limitation of meso-beta profile-correction is the influence of the orography: The profile measured in the vicinity of the radar is only in part representative for the precipitation over the Alps. This conclusion is supported by variogram analyses in the Alps (first article).

A cover of water or wet snow on the radome (the protection sphere around the radar antenna) may cause serious attenuation of the radar signal. In the third article we make a quantitative estimate of this type of attenuation by comparing average reflectivity in pre-defined precipitation systems before, during, and after rain on the radome. Seven events observed with the MeteoSwiss C-band radars are investigated. We found factors of attenuation ranging up to 5.4dB(Z) (two-way) for moderate rain at the radarsite. This cuts rain rates to less than half. Knowledge on the spatial continuity of precipitation, accurate precipitation measurements, and estimates of their quality are prerequisites for verifying and initialising weather prediction models with radar data. Future work will show to what extent radar data may help to improve precipitation forecasts and alerts in the Alps.
Zusammenfassung


a) die Verwendung von Radarreflektivitäts-Variogrammen zur Beschreibung der räumlichen Kontinuität des Niederschlags in den Alpen,

b) eine robuste Korrektur-Methode zur Verbesserung der Radar-Niederschlagsmessung in komplexer Orographie unter Benützung des vertikalen Profils der Radarreflektivität, und

c) eine Abschätzung der Abschwächung des Radarstrahls beim Durchdringen einer Wasserschicht auf dem Radome.

Im ersten Artikel verwenden wir Radarreflektivitäts-Variogramme, um die räumliche Kontinuität in verschiedenen Niederschlagstypen (stratiform versus konvektiv, Regen versus Schneefall) und in Abhängigkeit von der Lage relativ zu den Alpen (upstream, upslope, Alpenhauptkamm) zu quantifizieren. Die Variogramme liefern quantitative Antworten auf eine Reihe von Fragen, die mit der räumlichen Kontinuität des Niederschlags verknüpft sind, so zum Beispiel: Wie genau können wir den Flächenniederschlag aus Punktmessungen schätzen? Wie stark variiert die Niederschlagsrate in Abhängigkeit von der Distanz zwischen zwei Punkten (Vergleich von Messungen zweier Instrumente)? Welcher Zusammenhang besteht zwischen Auftreten von Konvektion und der Orographie? Wie gross ist die Ungenauigkeit einzelner Radarmessungen?


Eine Schicht aus Wasser oder nassem Schnee auf dem Radome (der Schutzhülle der Radaranenne) kann das Radarsignal stark abschwächen. Im dritten Artikel quantifizieren wir die Radome-Abschwächung während sieben Niederschlagsereignissen mittels Daten der drei MeteoSwiss C-Band Radars. Dazu wird die mittlere Reflektivität der Niederschlagssysteme vor, während, und nach dem Regen auf dem Radome verglichen. Abschwächungen bis zu 5.4dB(Z) (zweiweg) bei moderatem Regen wurden beobachtet. Die aus den Radarmessungen geschätzte Niederschlagsrate wurde in diesem Fall um mehr als die Hälfte unterschätzt.

Penso che la verità sia più grande di noi; gli alberi purtroppo ci nascondono ancora la foresta. Di sicuro è che io l’amo, e appena una minima particella d’essa mi si rivela, una minima certezza, mi sforzo di servirla. Dovrei essere cieca di mente per non accorgermi della limitatezza delle mie conoscenze: sono quelle di una ragazza che ha fatto l’università. Ma so che nei momenti in cui sento d’amare la verità in modo interamente disinteressato, allora anche il mio cervello s’arricchisce di nuove certezze. Penso che la verità ci è in gran parte sconosciuta perché gli uomini non l’amano abbastanza.

Ignazio Silone, Severina
Chapter 1

Introduction

1.1 Outline

This thesis consists of the following three articles


reproduced in the Chapters 2, 3 and 4, and of an introductory text putting the articles in the broader context of precipitation measurements by radar in the Alps.

1.2 The Challenge of using Radar for Precipitation Measurements in the Alps

Using weather radars for precipitation measurements in the Alps is a challenge. First, by lifting, channelling, blocking, by supplying moisture and by heating on sun-exposed slopes the mountains influence the flow and the stability of air masses from the synoptic down to the micro-scale, and thus play a key role in precipitation mechanisms in the Alps. The result is a complex picture of precipitation regimes and high variability on many scales. Second, the orography complicates precipitation measurements by radar because of severe ground clutter,
beam shielding, and difficult operating conditions on mountain sites. So, on the one hand
the Alps offer a variety of interesting research topics in the field of precipitation physics,
but on the other hand they pose particular difficulties for radar measurements. Note that
in the mountains also ground stations such as gauges are faced with problems: wind-induced
erors, lack of representativity in highly variable precipitation, heating needed to melt the snow,
snow burrying the instrument, data transfer, unreliable power supply and difficult access for
maintenance.

Quantitative applications of radar data are by far more demanding than qualitative ones. Before
discussing the problems of the former let us point out the benefit of the latter. A ground-based
radar is undoubtedly an excellent instrument for detecting and tracking precipitating regions
in real-time, even in a mountainous context. The main advantage with respect to a ground
station network is the high resolution in four dimensions. Weather radars give an overall picture
of the precipitation situation on the synoptic scale (composite images), monitor evolution and
path of precipitation on sub-synoptic scales, and provide detailed volume information on the
storm scale (e.g. by vertical cross sections of reflectivity and Doppler velocity). The great
help of weather radars for short-term predictions is undisputed. These goals are relatively easy
to achieve and are currently the main benefit from radars in the operational work of weather
services.

1.3 Problems and Solutions in an Alpine Context

Dominant sources of errors in quantitative precipitation estimates using radar data in the Alps
are ground clutter, shielding combined with the variations of the vertical profile of reflectivity,
partial beam-filling, attenuation in heavy rain, variations of the drop-size distribution (Z-R
relation), and hardware faults. Some sources are discussed in the following paragraphs, for
more details see Joss and Waldvogel (1990) and Joss and Germann (2000). Estimates of the
order of magnitude can be found in Joss (1998).

1.3.1 Clutter Elimination

Elimination of ground clutter from the main-lobe and the side-lobes, of anaprop, and of echoes
from airplanes, is a prerequisite for the use of radar data, both in a quantitative and qualitative
way. In mountainous regions where the radar sites are often on the top of mountains and the
lowest elevation angle is close to or even below the horizon, ground clutter poses a serious
problem.

In regions with frequent clutter the sensitivity for detecting precipitation is reduced. Both, the
reduced sensitivity and remaining clutter (not detected by the elimination algorithm), which vary from site to site and in time and space, limit the use of radar images.

A straightforward solution for clutter elimination is a static clutter map determined from a series of radar images in clear-sky conditions. The drawback of this solution is remaining clutter or the loss of valid precipitation measurements in regions with variable clutter (e.g. anaprop and side-lobe clutter). More sophisticated approaches combine all available information and use a dynamic clutter map. The algorithm proposed by Lee et al. (1995) is operational at the three radar sites of MeteoSwiss (Joss et al., 1998). It takes for each 83m raw gate a clutter/non-clutter decision using the radial velocity, the spectrum width, the minimum detectable signal (MDS), one-lag and two-lag signal fluctuations (statistical tests), the vertical gradient of reflectivity, as well as a continuously updated clutter map.

For the analyses presented in this thesis, in particular for the studies on the spatial continuity of Alpine precipitation (Chapter 2), only little remaining clutter can be accepted, but on the other hand holes caused by the clutter elimination must be kept to a minimum, and the complete three-dimensional reflectivity information is needed. Therefore, we put further effort to improve the operational clutter elimination. The results have been presented at the 24th EGS General Assembly in 1999 (abstract see Germann and Joss, 1999a). A modified version of the clutter decision-tree by Lee et al. (1995) was implemented on all three radar sites in summer 1999. The unmodified and the modified decision-tree are hereafter referred to as the old and the new elimination algorithm respectively.

In the remaining part of this section we briefly present and explain the modifications made in the operational algorithm. Taking a look at Lee et al. (1995), Joss et al. (1998) or Joss and Lee (1995) to understand the decision-tree is recommended.

To improve the clutter elimination we need to know why the remaining clutter is not detected by the decision-tree and where it occurs.

In Switzerland, the lowest elevation scan (-0.3deg) contains 23% cluttered pixels, making up for more than 70% of total clutter of the volume scan (consisting of 20 elevation angles). These estimates are obtained on La Dôle during fine weather by switching off the clutter elimination, then counting all pixels (1km x 1deg) with reflectivity larger than 13dBZ. The numbers are averaged over half-hour periods and, to some extent, may depend on the weather situation (refraction). The experiments have been repeated using the old and the new clutter elimination algorithm. Table 1.1 lists the counts. The lowest two elevations together (-0.3deg and 0.5deg) contain more than 90% \( ((10927+3285)/15503 = 92\%) \). Similar results have been obtained for the Monte Lema and Albis radar.

After the old elimination the remaining clutter has mainly non-zero velocity and preferably occurs in the vicinity of highly coherent zero-velocity clutter. This hypothesis is based on the following: \( N_{all} \) of Table 1.1 is the number of all cluttered pixels, while \( N_{mov} \) indicates the
Table 1.1: Frequency of clutter of La Dôle radar using no, old or new elimination. Clutter is identified during fine weather counting all pixels with reflectivity larger than 13 dBZ. Except for the group 'Whole range', we consider data within a range of 130 km from the radarsite. For each elevation and elimination mode we determined the number of all cluttered pixels ($N_{all}$), as well as those with non-zero velocity ($N_{mov}$). With non-zero we mean velocities larger than $0.125 \times$ the Nyquist velocity (at the lowest four elevation angles: unsigned radial velocity $>1\text{ms}^{-1}$). Doppler spectra are shown in Fig. 1.1. The total number of considered pixels is indicated in parenthesis after the elevation angle. See text for a discussion.

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<th>New</th>
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<td>$N_{mov}$</td>
<td>$N_{all}$</td>
</tr>
<tr>
<td>No</td>
<td>10927</td>
<td>114</td>
<td>3285</td>
</tr>
<tr>
<td>Old</td>
<td>613</td>
<td>584</td>
<td>203</td>
</tr>
<tr>
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<td>40</td>
<td>16</td>
<td>61</td>
</tr>
<tr>
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<td>$N_{mov}$</td>
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<tr>
<td>New</td>
<td>21</td>
<td>18</td>
<td>43</td>
</tr>
<tr>
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<td>3.8</td>
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<tr>
<td>New</td>
<td>402</td>
<td>279</td>
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\(^a\)Whole volume up to a range of 230 km.
Figure 1.1: Frequency distribution of Doppler velocity of clutter. Data is from two 2-hour periods of La Dôle at -0.3deg elevation and at ranges up to 130km. On average, when using no elimination 11060 (integral of curve of upper plot) of totally 46800 (130×360) pixels are cluttered. This corresponds to 24%. The small difference between 11060 (this figure) and 10927 of Table 1.1 is explained by the different periods. Upper plot: All clutter (no elimination). Lower plot (different ordinate!): All clutter versus clutter remaining after old elimination.

amount of clutter with non-zero Doppler velocity. With non-zero we mean velocities larger than 0.125 times the Nyquist velocity (at the lowest four elevation angles: unsigned radial velocity >1ms⁻¹). Clutter remaining with the old elimination algorithm has mainly non-zero velocity (946/990 = 96%). How can \( N_{\text{mov}} \) increase when using the old instead of no elimination? For each 1km gate we have data of twelve 83m raw gates. For the Doppler product, which is used to determine \( N_{\text{mov}} \), only the velocity of the 83m raw gate with the maximum reflectivity is taken. When switching off the clutter elimination, the velocity measurements of the Doppler product are from strong, highly coherent, zero-velocity clutter (342/15503 = 2.2%). When switching on the old elimination this type of clutter is easily detected by the statistical tests based on the signal fluctuations. That is why zero-velocity clutter mainly disappears with the old elimination. But from \( N_{\text{mov}} \) of the low elevation angles we conclude that zero-velocity
clutter is then in part replaced by neighbouring non-zero velocity clutter. The same is revealed by the broadening of the Doppler spectrum in the lower plot of Fig. 1.1. We conclude that clutter with non-zero Doppler velocity remaining after the old elimination preferably occurs in the vicinity of highly coherent zero-velocity clutter. Non-zero velocity clutter may be explained by antenna movement or simply by chance, caused by the big number of range gates (even with a small probability of occurrence some of the clutter signals have non-zero velocity).

To eliminate this type of clutter a neighbour-test is added in the decision-tree of the new algorithm: Gates with non-zero velocity only pass the clutter elimination algorithm if the adjacent gates have low clutter map entries. Adding the neighbour-test allows us to make the thresholds of the statistical tests less rigorous, thus eliminating fewer valid precipitation signals. The neighbour-test is active at the lowest two elevation angles only.

When using a dynamic clutter map particular attention must be paid to dead-ends and positive feedbacks. In our case, a continuous clutter map leak was needed to avoid a no-go situation for gates with zero Doppler velocity. In the old elimination algorithm, for a gate in a precipitation area (reflectivity above MDS) with zero Doppler velocity there was no way to decrease the clutter map entry once the gate was flagged as clutter (cf. decision-tree of Lee et al., 1995). This means a no-go as long as reflectivity is above MDS and the velocity is zero. Note that zero-velocity areas can be large and stationary (weather situation with weak winds, weak wind in the planetary boundary layer, wind perpendicular to radar-beam).

The last three numbers of Table 1.1 summarise the effectiveness of clutter elimination: Without elimination the radar volume contains 3% clutter ($19476/697320 = 3\%$), with the old elimination 0.2%, and with the new one 0.07%. Considering the lowest elevation only we obtain 23%, 1.3% and 0.09% respectively.

In addition to the neighbour-test and the clutter map leak a speckle filter has been added.

1.3.2 Profile-Correction

For three reasons in an Alpine country precipitation is often only visible at higher elevations: The radar is located on a mountain top, and the ridges as well as earth curvature hide the lower part of the atmosphere (Fig. 1.2). This combined with the systematic vertical decrease of reflectivity, or more precisely, the vertical profile of reflectivity, leads to errors between $-10\, \text{dB}$ and $+5\, \text{dB}$ (Joss, 1998). From the long-term average profile of Joss and Pittini (1991a), for measurements at 4km above sea level, we deduce an average underestimation with respect to the ground-level rate of $-6\, \text{dB}(\text{R})$ (factor of 4). Low-level precipitation, such as the one typically caused by a retour-d’est configuration in Ticino (private communication of the forecaster-team of MeteoSwiss, Locarno-Monti, see also Kappenberger and Kerkmann, 1997), may be completely missed by the radar.
Even in the absence of mountains and earth curvature the vertical variation of reflectivity would cause systematic errors. At a distance of 150km a one-degree beam has a vertical extent of 2.6km. The lower end of the beam may be in the rain but the upper part is often well above the melting layer in the snow where reflectivity is usually weak. Because of the systematic vertical decrease of reflectivity, the radar measurement, being an average over the pulse volume, on average underestimates the rate close to the ground. The pulse volume is proportional to the square of the distance (Fig. 1.4). Thus, the error increases with range. At close ranges echoes from the melting layer occasionally lead to overestimates, though.

A measurement from aloft is better than no measurement at all, and for qualitative use often sufficient, provided that precipitation reaches up to the height of the lowest radar measurement. For quantitative use, however, in regions where the lower part of the precipitation is not visible, measurements from aloft must be corrected for the vertical change of reflectivity (Chapter 3). First of all we need a scan strategy that provides measurements at several heights, such that we have higher-level data where the lower levels are obscured. The scan strategy is illustrated in Fig. 1.2 and briefly described in Section 1.4. Second, the reflectivity profile is strongly variable in time and space, therefore, the decisive point of profile-correction schemes is the estimation of a representative profile. Such an estimation procedure may use profiles measured in well-visible regions, e.g. in the vicinity of the radar (Joss and Lee, 1995), profile climatology (Joss and Pittini, 1991a), the four-dimensional information of reflectivity aloft plus knowledge on its correlation with the profile (Smyth and Illingworth, 1998b; Germann and Joss, 1999b), as well as other data sources such as gauges (Collier et al., 1983) or models (Kitchen et al., 1994).
1.3.3 Velocity Aliasing, Inhomogeneous Beam-filling and Attenuation

Compared to S-band C-band radars suffer less from ground clutter contamination, and, for a given antenna diameter (fixed antenna costs), have smaller beam-widths (better gain). Both are important factors in a mountainous region. C-band wavelengths are, on the other hand, more attenuated and have smaller Nyquist intervals.

(a) Velocity Aliasing

As a result of a small Nyquist interval, the radial velocity is often folded, once or several times. For real-time dual-Doppler wind field retrieval (Chong et al., 2000) during the MAP field experiment James and Houze Jr. (2000) developed a dealing a algorithm adapted to the low Nyquist velocity of Monte Lema data (8.25, 11.0 and 16.5ms⁻¹, see Table 1.2). To dealias velocity-azimuth-displays for the operational VAD wind profile product we use a first-guess wind estimated from the azimuthal derivative of the radial velocity (Germann, 1999b; Tabary et al., 2000).

(b) Inhomogeneous Beam-filling

Errors related to systematic vertical variations of reflectivity are addressed in Section 1.3.2. There, we found in shielded regions and at long ranges, on average the ground-level precipitation rate to be seriously underestimated. This underestimation occurs in spite of a slight overestimation of the rain rate that results from any inhomogeneity in the pulse volume when averaging linear reflectivity (this is what intrinsically happens in the pulse volume when signals of several scatterers with arbitrary phase are added). A correction would require an estimate of the spatial continuity of reflectivity (Chapter 2).

(c) Attenuation

The attenuation of a well designed, dry radome is small and constant (below 0.5dB). Attenuation caused by atmospheric gases is also quite constant (1.5dB per 100km). Both can easily be corrected in the radar equation. Serious signal attenuation may occur while propagating through a wet radome (Chapter 4), in hail, in heavy rain or in melting snow.

Many attempts have been made to correct attenuation errors. But the correction easily becomes unstable (Hitschfeld and Bordan, 1954), e.g. because of positive feedbacks in the algorithm (gate-by-gate iterative methods), and we must take care not to make things worse when correcting. To overcome this problem recent attempts make use of polarization diversity (e.g. Smyth and Illingworth, 1998a).
Estimating precipitation rate by means of KDP measurements (range derivative of differential phase) is free of attenuation problems. A recent example of combining KDP with reflectivity for quantitative estimates of rainfall can be found in Testud et al. (2000). For a review of polarimetry for weather surveillance radars see Zrnic and Ryzhkov (1999). The use of the KDP method in an Alpine context, however, is limited, because in large regions the lower part of the precipitation is shielded (see Figure 3.1.2 in Joss and Waldvogel, 1990) and the method does not work with data from the snow aloft.

1.3.4 Adjustment and Calibration

If enough care and effort is taken for maintenance and calibration, including automatic quality checks, errors caused by system instabilities and hardware faults can be neglected (Della Bruna et al., 1995).

For such a system we can determine the radar constant by comparing radar and rain gauge data, and thus adjust the radar precipitation estimates to ground truth. Because of different sampling volumes, and space and time lags we must take care when comparing radar and gauge measurements. The lack of representativity can be overcome by averaging over an appropriate space-time frame. The decisive point is choosing the right scale. In the literature we find methods ranging from real-time correction (Collier, 1986) to long-term adjustment on the seasonal scale (Joss and Lee, 1995). The article reproduced in Chapter 2 discusses the spatial continuity of Alpine precipitation and thus provides the basis for representativity studies. From radar data the variation of precipitation can not be determined down to the scale of gauges (diameter of 16cm). Nevertheless, we can derive some conclusions for radar-gauge adjustment: First, variograms like those presented in the article give a rough idea of the variability. Second, adjustment is not justified if the true difference between a radar and a gauge measurement is smaller than the expected difference read from the variogram at a lag of, e.g., 2km, roughly representing the time and space lag. Third, the article explains how to quantify the reduction of variance achieved by averaging in time and space. By comparing variograms of different regions, we finally find how the representativity of point measurements depends on the location in the Alps.

1.4 Data Basis

Data used in the studies of this thesis is from the operational C-band Doppler radar network of MeteoSwiss consisting of the three radars Monte Lema, La Dôle and Albis (Joss et al., 1998), as well as from the Mesoscale Alpine Programme (MAP) field experiment which took place in the European Alps in the period September-November 1999 (Bougeault et al., 2000).
### Table 1.2: Characteristics of MeteoSwiss radars.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Units</th>
<th>Monte Lema</th>
<th>La Dôle</th>
<th>Albris</th>
</tr>
</thead>
<tbody>
<tr>
<td>Frequency</td>
<td>GHz</td>
<td>5.44</td>
<td>5.43</td>
<td>5.45</td>
</tr>
<tr>
<td>Wave-length</td>
<td>cm</td>
<td>5.5</td>
<td>5.5</td>
<td>5.5</td>
</tr>
<tr>
<td>Peak power</td>
<td>kW</td>
<td>251</td>
<td>303</td>
<td>292</td>
</tr>
<tr>
<td>Loss factor (system)</td>
<td>dB</td>
<td>7.6</td>
<td>5.7</td>
<td>12.3</td>
</tr>
<tr>
<td>Antenna diameter</td>
<td>m</td>
<td>4.2</td>
<td>4.2</td>
<td>4.2</td>
</tr>
<tr>
<td>Gain</td>
<td>dB</td>
<td>44.7</td>
<td>44.7</td>
<td>44.7</td>
</tr>
<tr>
<td>3dB beam-width</td>
<td>deg</td>
<td>1.0</td>
<td>1.0</td>
<td>1.0</td>
</tr>
<tr>
<td>Pulse-length</td>
<td>µs</td>
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<td>0.5</td>
<td>0.5</td>
</tr>
<tr>
<td>Radial resolution</td>
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<td>83</td>
<td>83</td>
</tr>
<tr>
<td>Noise figure</td>
<td>dB</td>
<td>1.85</td>
<td>2.04</td>
<td>1.8</td>
</tr>
<tr>
<td>Band-width</td>
<td>MHz</td>
<td>2.0</td>
<td>2.11</td>
<td>2.08</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Transmitter</th>
<th>Magnetron</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of elevations</td>
<td>20 in 5min</td>
</tr>
<tr>
<td>PRF (depending on elevation)</td>
<td>Hz</td>
</tr>
<tr>
<td>RPM (depending on elevation)</td>
<td>min⁻¹</td>
</tr>
<tr>
<td>Pulses per 1km × 1deg pixel</td>
<td></td>
</tr>
<tr>
<td>Dynamic range logZ [I,Q]</td>
<td>dB</td>
</tr>
<tr>
<td>Maximum range</td>
<td>km</td>
</tr>
</tbody>
</table>

situation plan of the operational MeteoSwiss radars and the MAP research radars is given in Fig. 1.3.

An important property of weather radars is the range of various signals. Fig. 1.4 shows the received power above the noise level of the radar receiver of two point targets (mosquito, bird) and several volume scatterers versus distance from the radar. The radar characteristics can be found in Table 1.2. They determine the maximum range up to which a given rain rate can be detected, the range below which a strong signal (hail or main-lobe ground clutter) saturates the receiver, and the sensitivity for echoes from refractivity gradients (clear-air), or non-precipitating hydrometeors (cirrus, altostratus) which is crucial for calculating VAD wind profiles in the absence of precipitation.

Another decisive point of radar systems in mountainous regions is the choice of an appropriate scan strategy. At low elevation angles small elevation steps are needed to obtain measurements as low as possible above the Alps. Note that within the Alps the horizon which determines the height of the lowest possible measurement depends on both the azimuth and the range. Aiming at a high vertical resolution Sánchez-Diezma et al. (2000) propose elevation steps smaller than the 3dB beam-width. Measuring vertical reflectivity profiles in the vicinity of the radar, on the other hand, requires also high elevation angles. The pulse repetition frequency
1.4. Data Basis

Figure 1.3: Operational MeteoSwiss radars (Monte Lema, La Dôle and Albis) and the sites of the MAP research radars Ronsard C-band, S-Pol S-band, DOW X-band, OPRA S-band, LAPETH X-band, and KMRR K-band. OPRA and KMRR were set up at Locarno.
Figure 1.4: Sensitivity of Monte Lema radar: Received power of several scatterers for ranges between 100m and 230km. The quantities $Z$, $R$, $d$, $\sigma$ and $\eta$ are the reflectivity factor, the precipitation rate, the diameter, the back-scattering cross section, and the back-scattering cross section per unit volume respectively. To determine the receiver noise we assumed a temperature of 290K. Radar characteristics see Table 1.2.
1.5 Introducing the 3 Articles

The first of the three articles is entitled “Variograms of Radar Reflectivity to Describe the Spatial Continuity of Alpine Precipitation”. We aim to quantify how continuous, or how variable Alpine precipitation is. We use the word “continuity” rather than “variation” to place emphasis on the fact that close measurements are similar. Owing to the continuity we can compare measurements of different locations (e.g. weighted multiple regression of Gabella et al., 2000), extrapolate point observations to neighbouring areas (e.g. when producing rainfall maps from gauge network data), or estimate vertical profiles of reflectivity in partly shielded regions (profile-correction). Can variogram analysis quantify the spatial continuity of precipitation, produce a data base for representativity studies, solve the question where precipitation in the Southern Alps is most convective, and help to benefit from the high spatial and temporal resolution of radar data?

The second article discusses the problems around operational profile-correction in a mountainous region. Once ground clutter is reduced to an acceptable amount and the hardware is functioning properly, beam-shielding combined with the vertical decrease of reflectivity is the most important source of error for precipitation estimates by radar in the Alps. On the one hand, the vertical profile of reflectivity is highly variable in time and space, and it almost seems impossible to estimate the profile in partly shielded regions. On the other hand, the resulting errors are large. So any correction going in the right direction is to be done.

The studies on the spatial continuity of Alpine precipitation revealed systematic differences between the variograms in upstream regions (well-visible from Monte Lema), in upslope regions, and in valleys close to the main divide of the Alps (partly shielded). As a consequence, we can not expect the profiles observed in the vicinity of Monte Lema to be fully representative for the precipitation in the Alpine valleys. The question is how we can best estimate the profile and correct for it without introducing errors larger than the ones we want to correct. Working with profiles on the meso-beta scale the article presents a robust profile-correction scheme.

The last article is about signal attenuation caused by a water cover on the radome. We use reflectivity data of several rain events to determine the order of magnitude of radome attenuation of the MeteoSwiss C-band radars. The approach is empirical and requires, in contrast to theoretical simulations, no assumptions on the structure of the water cover. Reflectivity of
defined precipitation systems is analysed before, during, and after the rain at the radar site. The decrease of average reflectivity at the beginning of rain on the radome provides a rough estimate of radome attenuation. The difficult task is to separate radome attenuation from other changes of reflectivity (e.g. growth and decay of precipitation). The largest radome attenuation found in the analysed events is 5.4dB in terms of reflectivity, equivalent to more than a factor of 2 in terms of rain rate. The resulting underestimation may be acceptable for qualitative work but certainly must be considered when interpreting radar data quantitatively. Of course, the importance of radome attenuation depends on the frequency of precipitation at the radar site.
Chapter 2

Variograms of Radar Reflectivity to Describe the Spatial Continuity of Alpine Precipitation

2.1 Abstract

We use variograms of radar reflectivity as a summary statistic to describe the spatial continuity of Alpine precipitation on meso-gamma scales. First, we discuss how to obtain such variograms. Second, we want to find a set of typical variograms of Alpine precipitation. And third, we give some examples on how they can be used to tackle questions such as: What spatial variation of precipitation rate do we find in Alpine catchments? What difference do we have to expect between the measurements at two points separated by a given distance? To what accuracy can we estimate areal precipitation from point observations? Are there preferred regions for convection in Alpine precipitation?

Variograms are obtained using a method-of-moments estimator together with high-resolution polar reflectivity data of well-visible regions. Depending on the application we determined the variogram in terms of linear precipitation rate, logarithmic or linear reflectivity.

We found spatial continuity to significantly vary both in time and space in the various types of Alpine precipitation analysed so far. At a separation distance of 10km the expected difference of reflectivity ranges from 4dB(Z) (factor of 2.5 in stratiform rain or snow) to about 13dB(Z) (factor of 20 in a meso-scale convective system). In a 96-hour period of heavy rain in the Southern European Alps, maximum variation occurred in upslope regions (frequent convection), while close to the crest of the Alps the variation was relatively weak (persistent stratiform rain). The representativity of a point observation, which can be quantified given the variogram, therefore depends on both the time and the location within the Alps as well as on the integration time (integrated rainfall maps being less variable than instantaneous ones). For a 576km² basin and 40min average rain the fractional error of the basin precipitation
estimated by a gauge measurement ranges from 11% (variogram of stratiform autumn rain) to 65% (variogram of a meso-scale convective system). Next steps will extend the variogram analyses to a larger space-time domain towards a climatology of spatial continuity of Alpine precipitation.

2.2 Introduction

Statistical studies of spatial continuity of radar reflectivity, in particular by means of the variogram (or, alternatively, the autocovariance function), play a key role in many practical questions related to the natural variability of precipitation. We use the word continuity rather than variation to place emphasis on the fact that close measurements are similar. Owing to the continuity we can compare measurements of different locations (see e.g. weighted multiple regression of Gabella et al., 2000), extrapolate point observations to neighbouring areas (e.g. when producing rainfall maps from gauge network data), or estimate vertical profiles of reflectivity in partly shielded regions (profile-correction). An introduction to the analysis of spatial continuity can be found in Isaaks and Srivastava (1989). The first part of this paper presents the technique applied to obtain variograms of radar reflectivity in Alpine precipitation. In the second part we give examples how they can be used to tackle the following questions:

1. What spatial variation of precipitation rate do we find in the European Alps?
2. How does the variation decrease when averaging in time and space?
3. To what accuracy can we estimate the total water input in a catchment using a point measurement?
4. How can we compare precipitation measurements from two or more instruments in different locations? Does an observed difference lie within limits of meteorological variability, or is it significantly larger and must be interpreted as an instrumental difference?
5. Are there preferred regions for convection in Alpine precipitation? How does the frequency and the type of convection depend on the topography?
6. Can we use the nugget variance, i.e. the discontinuity observed at the origin of reflectivity semi-varigrams, to estimate the uncertainty of single radar measurements?

These questions, covering a wide range of meteorological and hydrological topics, are directly related to each other. The representativity of point measurements depends on the degree of spatial variation, which varies in time and space and as a function of the integration
time, as well as on the uncertainty of single measurements. With the uncertainty of single measurements we mean the difference between the measurement and the desired quantity, e.g. remaining scatter caused by signal fluctuations. If we find the spatial variation of precipitation to be much weaker close to the crest of the Alps than in upslope regions, we must consider this when interpreting gauge data in the Alps. Similarly, the comparison of measurements made in different locations (we ask whether two instruments agree) and the combination (radar-gauge adjustment) require the knowledge of the natural variability of precipitation at a specific location and time. These relations are the reason why we discuss the questions together.

Related work goes back to the beginning of quantitative radar meteorology. Austin and Houze (1970) combined reflectivity patterns with variance spectra of gauge data in order to investigate the relation between mesoscale precipitation areas and phenomena on larger and smaller scales. Zawadzki (1973) proposed an optical device to determine the space, as well as the Eulerian and Lagrangian time autocorrelation function (ACF) of radar images. For the analysed storm he found that the Taylor hypothesis (see e.g. Stull, 1993) holds up to periods of 40 min. If so, time and space can be exchanged, and we can compare, e.g., spatial variation in images of volume-scanning radars with time variation of data collected with vertical-pointing radars, the advantage of the latter being the high resolution in time and the vertical direction.

One problem of radar-gauge adjustment is the scatter introduced by the differences between the sample volumes. Different approaches have been proposed to quantify this effect. While Zawadzki (1975) takes ACFs obtained from gauge data of a 10 year period, Kitchen and Blackall (1992) determine the difference between point and areal values directly from rainfall maps drawn by eye on the basis of high-resolution gauge data. Ciach and Krajewski (1999) assume that radar-gauge differences can be partitioned into the error of the radar estimate and the lack of representativity of the gauge. To estimate the two components they propose a procedure that requires the ACF at scales below the size of the radar pixel. The nugget parameter of the exponential function used to model the ACF is estimated from data of five gauges a few km apart. A similar problem arises when verifying satellite-derived rainfall estimates with gauge data. Flitcroft et al. (1989) established a regression model which calculates the correction factor relating point measurements to areal averages. The underlying data is from a dense gauge network experiment.

The representativity of a point observation for area-mean values is a function of the spatial continuity. Ripley (1981) derives equations for the error variance for several standard schemes of spatial sampling given the ACF of the regionalised variable. Rodriguez-Iturbe and Mejia (1974a) discuss three techniques to relate point rainfall of a certain level of probability to areal rainfall with the same level of probability. They approximate the space ACF needed for one of the techniques using exponential and Bessel-type functions (Rodriguez-Iturbe and Mejia,
Another group of applications of precipitation variograms, not included in the six questions listed above, is rainfall modelling (e.g. Waymire et al., 1984), and kriging, e.g. cokriging radar-rainfall and rain gauge data, as proposed by Krajewski (1987), or detrended kriging considering the elevation-dependence (Garen et al., 1994). Variograms and related functions also play an important role in space-time downscaling of rainfall and its evaluation. The downscaling technique proposed by Venugopal et al. (1999) uses quantities that describe the multiplicative variation of precipitation, and preserves both the temporal and spatial correlation structure of rainfall.

In this paper we use high-resolution reflectivity data of well-visible regions (not shielded, free from clutter) to determine the range of variograms in precipitation in the European Alps. We focus on scales ranging from 0.5 to tens of km and from 5 min to a few hours (meso-gamma). This leaves the way open for several practical applications. Because of the influence of the orography on precipitation physics we expect a complex picture of variograms in the Alps. Section 2.3 gives an introduction to the variogram analysis. Section 2.4 presents the technique used to calculate variograms of radar reflectivity, followed by a set of typical variograms of Alpine precipitation in Section 2.5. In Section 2.6 we discuss the representativity of point measurements, quantify the effect of spatial averaging (change of support), and briefly explain how variograms have been used for an instrument-intercomparison experiment. Section 2.7 shows how variograms can be used in the context of orographic precipitation: We look for the regions of maximum orographic triggering of convection, given 96 hours of rain in the Lago Maggiore region (Southern Alps). In Section 2.8, finally, we estimate the nugget variance by extrapolating the semi-variogram to zero-lag and give a meteorological interpretation for it.

Gauges are scarce in the Alps, and suffer from site-specific attributes such as wind-induced errors (see e.g. Sevruk, 1989), losses caused by the wetting of the instrument or evaporation, and malfunction of the instrument. Weather radars, on the other hand, provide a high resolution in space and time. If the regions are carefully selected, contamination by measurement errors can be neglected. To what extent the integration of gauge data into the variogram analyses discussed in this paper will add new information, has to be shown in future work.

Here, data is from the operational C-band radar network of MeteoSwiss (Joss et al., 1998), and from the Mesoscale Alpine Programme field experiment (MAP, Bougeault et al., 2000). Problems and solutions when applying weather radars in an Alpine context are discussed e.g. in Joss and Germann (2000).
2.3 Spatial Continuity

Spatial and temporal continuity are intrinsic in meteorological data. Measurements from neighbouring stations tend to be more similar than those further apart. Even strongly variable properties, as e.g. rainrate in convective showers, show some spatial continuity. In general, correlation occurs whenever the phenomenon exhibits frequencies lower than the sampling frequency. We intuitively use this when planning a measurement campaign or designing a network of ground stations. For synoptic meteorology in a flat country there is no use of measuring absolute air pressure on a grid with one-kilometer meshes, because there will be almost no variation on this scale. A 20km mesh is likely to be sufficient.

Describing and interpreting the spatial information of earth science data is the main goal of geostatistics, initiated by the *Theory of Regionalised Variables* by Matheron (1962a, 1963) and his colleagues at the French mining school, and Krige (1951) after whom one of the most powerful interpolation techniques was named (Cressie, 1990). Matheron and Krige were both looking for statistical tools to improve ore-deposit evaluation using point measurements. Some terms, as e.g. the nugget effect, still refer to the mining roots of geostatistics.

Matheron (1962a) introduced the term *regionalised variable* (variable régionalisée) to underline the spatial aspect of earth science data. In Matheron (1963) he writes: “A regionalised variable is, sensu stricto, an actual function taking a definite value in each point of space.” In contrast to a random variable, its variation is, to some extent, correlated. On average the difference between samples increases with increasing separation distance. Extrapolation becomes straightforward if we can describe or approximate the variation by a mathematical law, e.g. that of the power distribution of the radar beam within the pulse volume, or the solar radiation at the earth’s surface given the terrain model. Usually there is no strict law (or we just do not know it). Then spatial continuity can be understood as a tendency or a probability. For neighbouring samples similar values are more probable than large differences, even if there is no strict relationship. Spatial variation is also a function of the scale (champ géométrique) and the support (support géométrique) of the data. The first term describes the extent of the whole region in which data has been collected. The second relates to the size, shape and orientation of the volume of each measurement (pulse volume in radar meteorology). Increasing the scale and decreasing the measurement volume both usually result in increasing variance since additional variation is taken into account.

2.3.1 The Variogram

The *variogram* is a tool to quantify spatial continuity of regionalised variables. It goes back to Matheron (1962a) and is defined as the variance of the difference between two values as a
Figure 2.1: Variograms of stratiform precipitation on 30 September 1998, 1000-1100UTC. Data is from region Maggiore, see Fig. 2.12. The variogram has been calculated for single images ($T_{ave} = 1$), and 10, 20, 40 and 60min average reflectivity ($T_{ave} = 2, 4, 8, 12$). The lower bundle of curves (thick lines) are the variograms, the upper show the number of data pairs involved. The diamond indicates twice the semi-variance ($2 \times 0.27dB^2(Z) = 0.54dB^2(Z)$) obtained for the h-scatterplot depicted in Fig. 2.2. The triangles mark the ones from the Figs. 2.3, 2.4, 2.5 and 2.6.

Function of the separation lag vector $h$

$$2\gamma(h) = \text{Var}\{Z(x) - Z(x + h)\} \quad (2.1)$$

where $2\gamma(h)$ is the variogram of the regionalised variable $Z(x)$ with actual values $z(x)$, and $\text{Var}\{}$ is the variance operator. It estimates the average squared difference between two measurements separated by a lag, or, more generally, tells us what spatial variation we have to expect for a given scale and support (example see Fig. 2.1).

The definition of the variogram assumes that $Z$ is intrinsically stationary which means that:

First, the expectation of $Z$ is constant throughout the whole area (stationarity in the mean)

$$\mathbb{E}\{Z(x)\} = \mu \quad (2.2)$$

where $\mu$ is the population mean, and $\mathbb{E}\{}$ the expectation operator. And second, $2\gamma(h)$ depends on the lag $h$ only and not on the position $x$, that is, for any subsample $V'$ of the whole region $V$

$$2\gamma'(h) = 2\gamma(h) = \text{Var}\{Z(x) - Z(x + h)\} \quad (2.3)$$
2.3. Spatial Continuity

If the region \( V \) is selected such that Eq. (2.3) is far from being fulfilled, Eq. (2.1) estimates a non-existent model parameter and the result depends on the position of the measurements and is only in part representative for the rest of the region. The variogram determined from a set of gauges all located in valleys is unlikely to be a perfect estimate of that in the mountains in between. With almost evenly spaced radar data it is less critical: If Eq. (2.3) is not valid we just obtain the average variogram of the region.

An alternate expression for the variogram can be derived by expanding the right side of Eq. (2.1)

\[
2\gamma(h) = \mathbb{E}\left\{ (Z(x) - Z(x + h))^2 \right\} + [\mathbb{E}\{Z(x) - Z(x - h)\}]^2
\]

From Eq. (2.2) follows that the last term vanishes and we obtain

\[
2\gamma(h) = \mathbb{E}\left\{ (Z(x) - Z(x + h))^2 \right\}
\]

What happens when the first assumption, the stationarity in the mean (Eq. (2.2)), is not satisfied? Then, the right side of Eq. (2.4) equals to the sum of the variogram (stochastic difference between two points) and the squared difference of the means (systematic difference). Whether this is desired or not depends on the application.

We can determine the variogram for different directions allowing to find whether spatial continuity shows preferred axes (anisotropy). Often anisotropy is weak or of secondary interest and an omnidirectional variogram \( 2\gamma(|h|) \) is all we need. The variogram is called isotropic if it only depends on the distance and not on the direction of the lag, that is \( 2\gamma(h) = 2\gamma(|h|) \).

Obviously, \( 2\gamma(h) = 2\gamma(-h) \), and \( 2\gamma(h) \) is always non-negative. Of particular interest are the values of the variogram as the lag distance approaches zero. By definition, \( 2\gamma(0) \) is zero and one might expect that as \( |h| \rightarrow 0 \) the variogram also tends to zero. However, in reality, we can often observe a discontinuity at the origin and \( \gamma(h) \rightarrow c_0 \) (non-zero) as \( |h| \rightarrow 0 \). This effect is called nugget effect, \( c_0 \) is the nugget variance (Matheron, 1963; Cressie, 1993). Two factors may be responsible for the observed discontinuity: micro-scale variability and measurement error. To illustrate the first let us assume you have found a gold nugget. However close to the nugget you go, the probability of finding another one will never rise to 1! In fact, it will usually be rather low. The measurement error includes instrumental problems causing a difference between the true and the measured value. Examples from weather radar systems are hardware deficiencies, signal fluctuations, or the system recovering after transmitting, which contaminates the measurements at close ranges. The separation of the two factors contributing to the nugget effect is not straightforward unless one of them can be neglected.
2.3.2 Covariogram and Correlogram

The definition of the autocovariance function \( C'(h) \) (covariogram) and the autocorrelation function \( \rho(h) \) (correlogram) requires similar assumptions as made for the variogram, see Eq. (2.2) and Eq. (2.3): First, stationarity in the mean, and second, for any subsample \( V' \) of the whole region \( V \)

\[
C'(h) = C(h) = \text{Cov} \{ Z(x), Z(x+h) \}
\]

(2.5)

where \( \text{Cov} \{ \} \) is the covariance operator. This assumption is stronger than Eq. (2.3), because here, in addition to Eq. (2.3), the population variance must be finite and constant (second-order stationarity). If Eq. (2.2) and Eq. (2.5) hold, the semi-variogram \( \gamma(h) \), the covariogram \( C(h) \) and the correlogram \( \rho(h) \) are related as follows

\[
\gamma(h) = \sigma^2 - C(h) = \sigma^2 (1 - \rho(h))
\]

(2.6)

where \( \sigma^2 \) is the population variance. According to Eq. (2.6) at large lags \( \gamma(h) \) tends to the population variance \( \sigma^2 = \text{Var} \{ Z \} \).

In practice the variance is often ill-defined because the analysed data set is just a sample of the population. Extending the dataset usually results in a larger variance. While the lower end (small lags) of the covariogram is directly affected, that of the variogram remains unchanged. An important advantage of the variogram is the fact that in order to calculate it, no assumption about the population variance is required. This is why we use the variogram rather than the more common autocovariance function.

2.3.3 Additivity

Let \( Z(p) \) be a regionalised variable measured in the scale \( A \) with support \( p \). The variance of any new variable derived by averaging \( Z(p) \) over support \( a \) (with \( p < a < A \)) is given by the difference of the variances of \( p \) inside \( A \) and \( p \) inside \( a \)

\[
\sigma^2(a, A) = \sigma^2(p, A) - \sigma^2(p, a)
\]

(2.7)

Equation (2.7) is known as Krige's formula (Matheron, 1963). Provided that \( Z(p) \) is intrinsically stationary (Eqs. (2.2) and (2.3)), \( \sigma^2(p, a) \) only depends on \( \gamma(h) \) of \( Z(p) \) inside \( a \)

\[
\sigma^2(p, a) = \frac{1}{a^2} \int_a \int_a \gamma(h)dx_1dx_2
\]

(2.8)

and not on the number of samples of support \( p \) within \( a \) as for a random variable. The vector \( h \) is the lag between the two area elements \( x_1 \) and \( x_2 \). Equation (2.8) defines the within-block
variance needed to estimate the kriging variance in block-kriging (Webster and Oliver, 1990). Substituting Eq. (2.8) in (2.7) yields

\[ \sigma^2(a, A) = \sigma^2(p, A) - \frac{1}{a^2} \int_a^a \int_a^a \gamma(h) dx_1 dx_2 \]  

which we will use in Section 2.6 to determine the effect of spatial averaging. By combining Eqs. (2.9) and (2.6) we obtain

\[ \sigma^2(a, A) = \sigma^2(p, A) \frac{1}{a^2} \int_a^a \int_a^a \rho(h) dx_1 dx_2 \]  

Thus, the expected autocorrelation between two randomly chosen points in a block of size \( a \) is exactly the ratio between the variances of \( Z(p) \) averaged over blocks of size \( a \) and \( Z(p) \). Note that the integral term never becomes zero because \( \rho(0) = 1 \). Rodriguez-Iiturbe and Mejia (1974a) propose to use the square root of the integral term

\[ \left( \frac{1}{a^2} \int_a^a \int_a^a \rho(h) dx_1 dx_2 \right)^{1/2} \]  

as a correction factor to transform point rainfall rate with a given return period (and variance \( \sigma^2(p, A) \)) to areal rainfall rate with the same return period (and variance \( \sigma^2(a, A) \)).

A more general discussion on change of support can be found in Isaaks and Srivastava (1989).

### 2.3.4 The Semi-variogram and h-scatterplots

Why is \( \gamma(h) \), known as the semi-variogram, defined as half the variance of \( Z(x) - Z(x + h) \)? Where does the factor 2 in Eq. (2.1) come from?

The covariogram (Section 2.3.2) is basic to the minimization of the kriging variance (Isaaks and Srivastava, 1989; Matheron, 1962b). This is because the variance of a weighted linear combination of random variables distributed in space is a function of the covariogram and the weights (Eq. (2.16)). Provided the population variance is constant and finite, the covariogram equals to the difference between the population variance and the semi-variogram (see Eq. (2.6)), and one can be replaced by the other. Traditionally the semi-variogram rather than the covariogram is estimated and modelled for kriging.

There is another argument for the factor 2 in Eq. (2.1): \( \gamma(h) \), the semi-variance at lag \( h \), can be interpreted as the normalised moment of inertia about the 45 degree-line of an h-scatterplot showing all pairs of data separated by lag \( h \) (Fig. 2.2). It is given by

\[ I(h) = \frac{1}{2N(h)} \sum_{i=1}^{N(h)} [z(x_i) - z(x_i + h)]^2 \]  

(2.12)
Thus, the semi-variogram is a vector of normalised moments of inertia, each element being a summary statistic of an h-scatterplot. Note the analogy between $I(h)$ and the moment of inertia about an axis of rotation in mechanics.

An example of an h-scatterplot is depicted in Fig. 2.2. Data is from Monte Lema radar at 0.5deg elevation, 30 September 1998, 1000-1100UTC. A well-visible region of 914km$^2$ to the south of the radar site with no clutter-problems, i.e. no remaining clutter, has been chosen: Azimuth between 135 and 251deg at ranges between 9.5 and 31.5km (region Maggiore, Fig. 2.12). The reflectivity discretisation of 0.25dBZ results from averaging 12 PPIs (one hour) originally having 3dBZ class-width. For the depicted data pairs we obtain $I(1148m) = 0.27dB^2(Z)$. This value, multiplied by 2, appears in Fig. 2.1 as a single point of the 60min average reflectivity variogram (see diamond). As the lag increases so does the fatness of the cloud (compare Fig. 2.2 with 2.3), as well as the variogram (see Fig. 2.1).

The cloud of points in Fig. 2.3 looks asymmetric, more points lie below the line than above. The reason for this, however, is of no interest. Each data pair separated by the given lag is only considered and plotted once. Whether it appears to the upper-left or the lower-right side of the 45 degree-line is just a question of order of data processing. The asymmetry is irrelevant for our studies.

### 2.4 Estimating the Variogram of Radar Reflectivity

According to the definition given in Eq. (2.1) the variogram of radar reflectivity is the variance of the difference between the reflectivity at two locations separated by the lag $h$. To estimate the variogram from a given data set Matheron (1962a) proposes

\[
\hat{\gamma}(h) \equiv \frac{1}{N(h)} \sum_{i=1}^{N(h)} [z(x_i) - z(x_i + h)]^2
\]  

which is a biasfree estimator of Eq. (2.4). As the sample mean is not needed we devide by $N(h)$ and not by $N(h) - 1$.

Whether we use linear precipitation rate (mm/h), logarithmic (dBZ) or linear reflectivity (mm$^6$m$^{-3}$) for variogram analysis depends on the application, the variogram estimator and the characteristics of the underlying data.

The classical variogram estimator of Matheron (1962a) using second moments (Eq. (2.13)) shows little robustness: Non-normality (e.g. skewness and heavy tails) and outliers (e.g. remaining clutter) seriously affect the result. Cressie and Hawkins (1980) and Hawkins and Cressie (1984) discuss the influence of non-normality and outliers, and propose a robust estimator instead of Eq. (2.13). A summary can be found in Cressie (1993). As an alternative
2.4. Estimating the Variogram of Radar Reflectivity

Figure 2.2: h-scatterplot of 60min average reflectivity for lag distance 1148m ± 0.4dB, 30 September 1998, region Maggiore. Number of used data pairs is 5266. For the semi-variance we obtain $0.27dB^2(Z)$.

Figure 2.3: h-scatterplot as in Fig. 2.2, lag distance 7244m ± 0.4dB.
Figure 2.4: $h$-scatterplot of raw reflectivity (no time averaging). Data pairs from the same lag and space-time window as in Fig. 2.2.

Figure 2.5: $h$-scatterplot of 10min average reflectivity (2 PPIs). Data pairs from the same lag and space-time window as in Fig. 2.2.
2.4. Estimating the Variogram of Radar Reflectivity

Estimating the Variogram of Radar Reflectivity

Reflectivity \( Z(x) \) [dBZ]

Figure 2.6: \( h \)-scatterplot of 20min average reflectivity (4 PPIs). Data pairs from the same lag and space-time window as in Fig. 2.2.

to robust estimators we may transform the data to normality, carefully select the data window and check for outliers, e.g. by means of \( h \)-scatterplots. For reflectivity the difference \( z(x_i) - z(x_i + h) \) is approximately lognormal, therefore, Eq. (2.13) is a robust variogram estimator if using logarithmic reflectivity. Variograms obtained this way describe the multiplicative character of variation. In Fig. 2.2 the fatness of the cloud, and thus the multiplicative variation of rain, is approximately independent of the intensity, which is an example of multiplicative behaviour of precipitation.

In Section 2.6.1, however, we need variograms of linear precipitation rate, the probability distribution of which is highly skewed. Then the poor robustness of Eq. (2.13) can be overcome as follows: 1) Select a representative data set such that the tails of the distribution are well-defined, and 2) no outliers are present. 3) Average in time to reduce the skewness of the data (central limit theorem).

2.4.1 Data Base and Processing

Why do we use polar data (PPIs with 1km \( \times \) 1deg \( \times \) 1deg resolution) for variogram analysis? Polar data is clutter-filtered and, at close ranges, provides high spatial resolution, required to estimate the variogram at short lags. Second, it has the same support as the original measurements. Resampling, interpolation and averaging required for the derivation of cartesian products, such as horizontal sections and vertical maximum projections, would affect the estimation of the variogram. Third, the knowledge of spatial variation of reflectivity helps to
design profile correction schemes, most of them are also based on polar data. The methods presented in recent papers work with measured, retrieved and idealised profiles determined and applied on various scales, ranging from climatological profiles (long-term averages) depending on the season or the rain type, to several types of meso-scale profiles (for an overview see e.g. Smyth and Illingworth, 1998b). As the shape of the profile is correlated with reflectivity (Fabry and Zawadzki, 1995) and its variation (Germann and Joss, 1999b), we suggest to use variograms to investigate the representativity of the profiles on these various scales, and, further, in order to check the validity of the assumption of spatial homogeneity of the profile made in Andrieu and Creutin (1995) and Vignal et al. (1999). Studies of this type, however, are beyond the scope of this paper.

We know that meteorological radars have both, sources of radial (azimuth-dependent) and concentric (range-dependent) error patterns: Beam-blocking is the most important source of the first type of error, while ground clutter, the dependence of the pulse volume on the distance, and the vertical change of reflectivity including bright-band effects belong to the second group. As a result, if variograms were calculated separately for different azimuths and distances, we would observe ‘polar’ anisotropy. The sources of errors listed above also affect omnidirectional variograms: Beam-blocking, mountain returns and the vertical change of reflectivity result in overestimating the variogram, because artificial variation of reflectivity is added to the meteorological one. Changing the pulse volume by varying the distance to the radar means changing the support, larger supports having smaller variance (see Krige’s formula and Section 2.6.2).

We can avoid these problems 1) by selecting an appropriate space-time window where the errors can be neglected, and, 2) by correcting for them. If, e.g., the height of the measurements varies too much (high elevation angles and/or large range intervals) polar data must be corrected for the vertical reflectivity profile before variogram estimation. Close to the radar at low elevation angles the vertical extent of the pulse volume and the variation of height within the space window are small. Furthermore, if the zero-degree isotherm is considerably higher than the selected data, the influence of the profile is negligible.

There are also meteorological reasons for anisotropy, e.g. parallel rain bands or alignment of cells. In these cases it might be interesting to look at variograms obtained for several azimuth intervals. Since we observed no dominant anisotropy we used omnidirectional variograms $2\gamma(h) = 2\gamma(|h|)$ with the advantage of having more data pairs for variogram estimation. In contrast to other statistical techniques, variogram analysis does not require data gap filling. The variogram is estimated on the basis of all valid, clutter-free and well-visible data pairs, missing data is ignored. Daily clutter maps and long-term rain totals provide information on the visibility of a region of interest (Joss and Lee, 1995).

We are interested in reflectivity variation within rain rather than at the borders. Therefore,
we restrict our analysis to regions with reflectivity $>13$dBZ and select space-time windows providing a minimum percentage of samples above this threshold (here 50%).

The distance between two measurements is determined from the centres of the pulse volumes. $2\hat{\gamma}(h)$ is calculated from all data pairs with lag distance $|h| \pm 0.4$dB ($\pm 10\%$). The class-width of 0.8dB has been selected such that each class is based on a sufficient number of samples, however small enough to analyse small-scale variability of reflectivity, our major interest. Since the radial resolution of the underlying data is 1km, at lags smaller than 1km $2\hat{\gamma}(h)$ is based on data pairs at constant range. Such small lags, for instance, are required to estimate the decrease of the variance resulting from averaging a few neighbouring pixels (see Section 2.6.2). As a reasonable maximum lag we propose one fourth of the extent of the space window.

2.4.2 Time Averaging

A common technique to reduce the stochastic part of any type of error is averaging in time. Its effect on reflectivity variation can be quantified by calculating variograms for different averaging intervals. Say we want to know the multiplicative variation of reflectivity at two points separated by a certain lag as a function of the averaging time. Figure 2.1 depicts the variograms of raw polar data, as well as 10, 20, 40 and 60min average reflectivity during stratiform rain in September 1998. Time resolution of Monte Lema data is 5min, thus 10, 20, 40 and 60min correspond to averaging 2, 4, 8 and 12 images.

If the data pairs were independent, the variance would decrease with $1/N$, where $N$ is the number of integrated images. Variation at small lags and between successive volume scans is of the same order of magnitude. Therefore, the difference of reflectivity measured at two locations is almost independent from the one in the previous scan. In fact, at lags below 1km the variograms in Fig. 2.1 approximately decrease with $1/N$. Note that this depends on the temporal and the spatial resolution of the underlying data. Rainfall is well known to have structure down to much smaller scales (Fabry, 1996). Data pairs sampled with a higher resolution in time, e.g., would be correlated and the variance would decrease less rapidly. As a consequence of the decrease with $1/N$ at small lags averaging a few pixels in time and/or space efficiently reduces the sample variance, and we obtain a better estimate for comparison with other point measurements. At larger lags correlation in time becomes important and time averaging is less efficient; $2\hat{\gamma}(10\text{km})$, e.g., decreases with $1/N^*$ where $N^* = N/(0.78+0.22N)$.

Again h-scatterplots provide deeper insight into the data and help to interpret the variogram. Using the same data and lag we calculated h-scatterplots with raw, as well as 10 and 20min average reflectivity (Figs. 2.4, 2.5 and 2.6). The semi-variance decreases from $2.61\text{dB}^2(Z)$ with raw data to $1.32\text{dB}^2(Z)$ (10min average) and $0.69\text{dB}^2(Z)$ (20min average). See also triangles in Fig. 2.1.
Figure 2.7: Typical variograms of Alpine precipitation: Stratiform Autumn (30 September 1998), Stratiform Winter (2 December 1997), Meso-scale Convective System (MCS) (19 August 1998). Data is from the region Maggiore, see Fig. 2.12. All variograms are calculated from raw reflectivity except for the MCS five-hour period, for which both raw and 60min average reflectivity has been used. The variogram Stratiform Autumn Rain has been discussed in detail in Fig. 2.1.

2.5 Variograms of Alpine Precipitation

Figure 2.7 depicts a set of typical variograms of Alpine precipitation. We selected a stratiform event from winter 1997, one from autumn 1998 and a meso-scale convective system (MCS) from summer 1998. Variograms of measurements in snow have been calculated for both the autumn and the winter case. For the autumn case we used 4.5deg elevation data of Monte Lema radar (1625m ASL) at 20km range (3.2km ASL, 500m above the melting layer), and a region with a small radial extent (2km ×116deg) to avoid errors caused by the vertical reflectivity profile (see Section 2.4.1). The other five variograms base on 0.5deg elevation data allowing a region with a much larger radial extent (region Maggiore, Fig. 2.12: 22km ×116deg, 1.8km ASL). In the winter case the zero-degree isotherm fell from 1.2km down to 900m and was well below the selected measurements at 0.5deg elevation. The one-hour period of the MCS is at the time of maximum convection while the five-hour period almost covers its whole life-cycle in the selected region.

The variograms differ by more than a factor of 10. At a separation distance of 10km the expected difference of reflectivity during the hour with maximum convection of the MCS is
Figure 2.8: Same as Fig. 2.7, but with linear abscissa. The triangle to the right marks twice the population variance of the data of 'Stratiform Autumn Rain 1 hour'.

of the order of 13dB(Z), on an average within the MCS 7.6dB(Z), in the stratiform rain 4.3dB(Z), and in the stratiform snow 3.8dB(Z). Note the small difference between the two variograms in freezing conditions.

2.6 Representativity of Point Measurements

In this section we 1) determine the accuracy of areal precipitation estimated from point measurements such as gauge data, 2) quantify the effect of spatial averaging using Krige's formula (Section 2.3.3), and 3) briefly communicate an idea how variograms may help to interpret measurements from instruments in different locations.

2.6.1 Mean Basin Precipitation

Let $G$ be the precipitation rate measured with a gauge over period $t$. Say this single measurement is used to estimate the rate $R$ averaged over a drainage basin of size $B$ and period $t$, assuming a constant rate in $B$. We now divide the basin in $N$ pixels of the same size as the support $p$ for which we determined the variogram (region Maggiore: 0.2km$^2$). Let $r_i$ with $i = 1, \ldots, N$ be the true rate of the $N$ pixels, the gauge lying in the pixel with $r_1$. Further let $\hat{r}_1$ be $r_1$ estimated from the gauge measurement using $\hat{r}_1 = G$ with error variance $s^2_{\hat{r}_1}$.
We assume that for the selected space-time window precipitation is intrinsically stationary (Section 2.3.1). The true areal precipitation rate is defined by

\[ R = \frac{1}{N} \sum_{i=1}^{N} r_i \]  

and estimated from \( \hat{R} = \hat{r}_1 \). To find out how accurate \( \hat{R} \) is, we use the expected squared difference \( s_{\hat{R}}^2 \) between \( R \) and \( \hat{R} \), which is equal to the variance of \( R - \hat{R} \) because \( \hat{R} \) is unbiased.

\[ s_{\hat{R}}^2 = \text{Var}\{R - \hat{R}\} \]

Using the equation for the variance of a weighted linear combination of random variables (see e.g. Isaaks and Srivastava, 1989, page 217)

\[ \text{Var}\left\{ \sum_{i=1}^{N} w_i X_i \right\} = \sum_{i=1}^{N} \sum_{j=1}^{N} w_i w_j \text{Cov}\{X_i, X_j\} \]

where \( w_i \) are the weights and \( X_i \) the random variables, we can write

\[ s_{\hat{R}}^2 = \text{Var}\{R - \hat{R}\} = \text{Var}\{R\} + \text{Var}\{\hat{R}\} - 2\text{Cov}\{R, \hat{R}\} \]  

The argument of the first term is itself a linear combination of random variables, therefore,

\[ \text{Var}\{R\} = \text{Var}\left\{ \frac{1}{N} \sum_{i=1}^{N} r_i \right\} = \frac{1}{N^2} \sum_{i=1}^{N} \sum_{j=1}^{N} \text{Cov}\{r_i, r_j\} \]

The second term of Eq. (2.17) is

\[ \text{Var}\{\hat{R}\} = \text{Var}\{\hat{r}_1\} = \text{Var}\{r_1\} + s_{\hat{r}_1}^2 \]

In case \( G \) is uncorrelated with \( R \) the last term of Eq. (2.17) becomes zero and \( s_{\hat{R}}^2 \) is simply the sum of Eqs. (2.18) and (2.19). A 10min measurement of a gauge during a thunderstorm, e.g., often exhibits poor correlation with the average rain in a larger area. However, usually
we select the period and the area such that the correlation is significant. Then $s_R^2$ decreases by the variance explained by $G$, that is, by the third term of Eq. (2.17)

$$2\text{Cov}\{R, \hat{R}\} = 2\text{Cov}\left\{\frac{1}{N} \sum_{i=1}^{N} r_i, \hat{r}_1\right\}$$

$$= \frac{2}{N} \sum_{i=1}^{N} \text{Cov}\{r_i, \hat{r}_1\}$$

$$= \left(\frac{2}{N} \sum_{i=1}^{N} \text{Cov}\{r_i, r_1\}\right) - 2s_{\hat{r}_1}^2$$

We can now rewrite Eq. (2.17) as follows

$$s_R^2 = \frac{1}{N^2} \sum_{i=1}^{N} \sum_{j=1}^{N} \text{Cov}\{r_i, r_j\}$$

$$+ \text{Var}\{r_1\} + 3s_{\hat{r}_1}^2$$

$$- \frac{2}{N} \sum_{i=1}^{N} \text{Cov}\{r_i, r_1\}$$

If the support $p$ is small and the integration time $t$ long, $s_{\hat{r}_1}^2$ can be neglected, and together with Eq. (2.15) we obtain the final expression for the expected squared error

$$s_R^2 = \frac{1}{N^2} \sum_{i=1}^{N} \sum_{j=1}^{N} \text{Cov}\{r_i, r_j\}$$

$$+ \text{Var}\{r_1\} - \frac{2}{N} \sum_{i=1}^{N} \text{Cov}\{r_i, r_1\}$$

or, alternatively, as a function of $\gamma(h)$ of $r(p)$

$$s_R^2 = \frac{2}{N} \sum_{i=1}^{N} \gamma(h_{i1})$$

$$- \frac{1}{N^2} \sum_{i=1}^{N} \sum_{j=1}^{N} \gamma(h_{ij})$$

where $h_{ij}$ is the separation vector between the two area elements $i$ and $j$. The first term is twice the expected value of the semi-variogram between the gauge and an arbitrary point in the drainage basin, and quantifies the representativity of the gauge for another point in the area. The second term is the expected value of the semi-variogram between two randomly chosen points in the basin, correcting for the fact that the variance of the basin precipitation decreases with increasing area. The difference between the two gives the desired quantity.
Now we use Eq. (2.23) together with variograms of linear rainrate of the two events 'Stratiform Autumn' and 'MCS' discussed in Section 2.5. Data is from the region Maggiore, Fig. 2.12. The rainrate has been averaged over 40min, a reasonable period for hydrological applications. Values smaller than 0.16mm/h have been omitted for variogram estimation because they are, for the most part, caused by clutter elimination and are not true zero-rain pixels. Thus, true zero-rain is also ignored, and the resulting $s_R^2$ estimates the lack of representativity of a non-zero point measurement for a non-zero rain area. In case of isolated thunderstorms zero-rain areas become relevant and must be considered. But then, the important question may be different from the one discussed here, e.g.: how accurate can we estimate the maximum local one-hour rain amount from a set of point measurements?

Long-term accumulations show that in the region Maggiore variations in the mean can be neglected compared to the variations in 40min average fields. Thus Eq. (2.2) is fulfilled, and the calculated variograms describe the spatial variation of precipitation as defined in Eq. (2.1). Any nugget variance caused by measurement error is undesirable and must be subtracted from the observed variogram. Table 2.1 lists the values of the variograms at a lag of 5km, as well as the mean and the variance of the underlying data.

We defined two rectangular basins with 576 and 144km² respectively (Fig. 2.9). The size of the large one reflects the density of fast-response gauges in the southern Swiss Alps. The expected error of the gauge-estimate of the average 40min precipitation in the basin depends on the precipitation type (variogram), on the size and the shape of the basin, and on the location of the gauge. Table 2.2 shows the results. Together with Table 2.1 we find: First, the fractional error ($s_R^2$/Mean) ranges from small (11%, variogram of stratiform autumn rain) to serious (65%, variogram of MCS rain 1hour). On average in the MCS (variogram of MCS five-hour period) the fractional error is 43%. Second, using the MCS variograms we obtain, as one would expect, a smaller error for the smaller basin. In the stratiform autumn rain it is the opposite. This because the first term of Eq. (2.23) increases less than the second when switching from the small to the large basin. In other words, in the stratiform autumn rain the uncertainty introduced by larger distances is less important than averaging over a larger area which reduces the overall variance. Third, discussing the representativity of a gauge measurement one usually bears in mind the first term of Eq. (2.23), but often misses the importance of the second.

Next steps will be to extend this analysis to larger integration times (e.g. daily precipitation), to consider a set of point measurement for area-mean estimates, and by means of daily clutter maps to distinguish between true zero-rain and zero-pixels that are caused by clutter elimination. The results will form an important basis of error analysis in climatological studies, such as the precipitation climatology of the European Alps by Frei and Schär (1998).
### Table 2.1: Statistics of variograms of linear precipitation rate.

<table>
<thead>
<tr>
<th>Event</th>
<th>Mean (mm/h)</th>
<th>Variance (mm/h)^2</th>
<th>2(\gamma(5\text{km})) (mm/h)^2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stratiform Autumn Rain</td>
<td>1.0</td>
<td>0.22</td>
<td>0.066</td>
</tr>
<tr>
<td>MCS Rain 5hours</td>
<td>1.9</td>
<td>1.1</td>
<td>1.1</td>
</tr>
<tr>
<td>MCS Rain 1 hour</td>
<td>5.5</td>
<td>33</td>
<td>26</td>
</tr>
</tbody>
</table>

### Table 2.2: Expected error \(s_R\) of gauge-estimate of 40min-average basin precipitation.

<table>
<thead>
<tr>
<th>Basin/Event</th>
<th>Equation (2.23)</th>
<th>(s_R) (mm/h)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1st term (mm/h)^2</td>
<td>2nd term (mm/h)^2</td>
</tr>
<tr>
<td><strong>576km^2 basin</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Stratiform Autumn Rain</td>
<td>0.16</td>
<td>0.14</td>
</tr>
<tr>
<td>MCS Rain 5 hours</td>
<td>1.8</td>
<td>1.1</td>
</tr>
<tr>
<td>MCS Rain 1 hour</td>
<td>59</td>
<td>46</td>
</tr>
<tr>
<td><strong>144km^2 basin</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Stratiform Autumn Rain</td>
<td>0.068</td>
<td>0.044</td>
</tr>
<tr>
<td>MCS Rain 5 hours</td>
<td>1.1</td>
<td>0.64</td>
</tr>
<tr>
<td>MCS Rain 1 hour</td>
<td>27</td>
<td>18</td>
</tr>
</tbody>
</table>

### Table 2.3: Reducing by spatial averaging the variance of 40min-average precipitation.

<table>
<thead>
<tr>
<th>Event</th>
<th>2nd term of Eq. (2.23) (mm/h)^2</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>16km^2</td>
</tr>
<tr>
<td>Stratiform Autumn Rain</td>
<td>0.017</td>
</tr>
<tr>
<td>MCS Rain 5 hours</td>
<td>0.31</td>
</tr>
<tr>
<td>MCS Rain 1 hour</td>
<td>6.8</td>
</tr>
</tbody>
</table>
2.6.2 Spatial Averaging

From Eq. (2.9) follows that the second term of Eq. (2.23) is exactly the amount of variance reduced by averaging $r(p)$ over the basin area $B$. Thus, the second term of Eq. (2.23) can be used to determine the effect of spatial averaging. Results for three square boxes and the same three variograms as above are listed in Table 2.3. Comparing these values with the variance of the underlying data given in Table 2.1 yields the relative reduction of variance. Averaging over the 16km² box, e.g., results in a reduction of the variance of 8, 28, and 21% respectively.

Another example of spatial averaging is the vectorial sum of the echoes of all scatterers in the pulse volume. We may ask about the influence of the increase of the pulse volume with range on the determination of the variogram. Within the region Maggiore, the size of a one-degree pixel ranges from $0.17 \times 1$km² (at 9.5km) to $0.55 \times 1$km² (at 31.5km). Measurements at long ranges, being an average over larger pulse volumes, have smaller meteorological variance. Both a theoretical and an experimental analysis show that this influence can be neglected: For the theoretical estimate of the reduction of the variance we again use Eq. (2.9). Here, variograms of linear reflectivity are needed because reflectivity is linearly averaged in the pulse volume. The nugget variance attributed to signal fluctuations must be subtracted before using $\gamma(h)$ in Eq. (2.9). Let $p$ and $a$ of Eq. (2.9) be the support at 9.5 and 31.5km respectively. In precipitation represented by the variogram of the MCS five-hour period, e.g., $\sigma^2(p, a)/\sigma^2(p, A)$ becomes 0.06, i.e. the variances of measurements of linear reflectivity at 9.5 and 31.5km would roughly differ by 6%. A second estimate of the influence of the size of the pulse volume is obtained with the following experiment: Calculating variograms of two regions at ranges between 9.5 and 20.5km and between 20.5 and 31.5km respectively, reveals small differences (<0.3dB²(Z)), much smaller than those found between the various precipitation types (tens of dB²(Z), see Fig. 2.7). To ensure that we observe the same precipitation in both regions, we select a 24 hour period of stratiform rain and define two equally sized regions on a line along echo movement.

Later on, in Section 2.7, we compare variograms of the same time period but of different
regions. Take, for instance, the regions Gridone and Vigezzo (Fig. 2.11) lying at 17 and 34km respectively. Using again Eq. (2.9) and the variogram of linear reflectivity of the MCS five-hour period, we obtain a decrease of the variance of approximately 7% which is again negligible.

### 2.6.3 Instrument Intercomparison

The last topic of this section is the comparison of measurements of two or more instruments in different locations. Here, we just briefly describe the idea. We want to know whether an observed difference lies within the limits of meteorological variability or is significantly larger and hence must be interpreted as an instrumental difference. The uncertainty introduced by the fact that two observations are from different locations and precipitation varies in space can be estimated by means of reflectivity variograms. This approach has been followed to interpret the differences between vertical reflectivity profiles of two volume scanning and two vertically pointing radars operated in the vicinity of Monte Lema during the intercomparison experiment preceding the special observing period of MAP (Bougeault et al., 2000).

### 2.7 Upstream and Upslope Variograms

Stratiform precipitation is characterised by a rather homogeneous intensity distribution and weak horizontal gradients. Convection, on the other hand, typically produces regions with high reflectivity and strong horizontal gradients reflecting the alternation of updrafts and downdrafts (Houze, 1997). Steiner et al. (1995) proposed a classification scheme based on horizontal reflectivity gradients to separate stratiform and convective rain. Here, we use reflectivity variograms of Alpine precipitation to find the regions with maximum orographic triggering of convection.

We compare variograms of regions in upstream and upslope conditions during heavy precipitation in the Southern Alps. Seven days in September 1999 (17-21 and 25-26 September, i.e. MAP IOP-02a, IOP-02b and IOP-03, Bougeault et al. (2000)) with 96 hours of rain have been selected. The gauge at Locarno-Monti registered a total amount of 394mm. Operational wind profiles of Monte Lema radar (Germann, 1999b) provide information on the main wind flow (Fig. 2.10). During the selected period at 2km ASL the wind blew from SSE with 5 to 20ms$^{-1}$, at 5.6km from SSW with 10 to 40ms$^{-1}$. Reflectivity profiles of Monte Lema and two vertically-pointing radars (OPRA S-band at Locarno-Monti and ETH X-band at Macugnaga, Fig. 2.11) as well as radiosonde data of Milano are used to determine the range of height of the melting layer (Fig. 2.15).

Figures 2.11 and 2.13 depict the selected regions and the corresponding variograms using
Figure 2.10: Direction of the meso-scale wind at two levels (2 and 5.6km ASL) during the 96 hours of September 1999 rain.

Monte Lema data. The inner three regions (Gridone, Valcolla and Campo dei Fiori) are used together with 1.5deg elevation data, the outer three with 0.5deg data (Fig. 2.15: Variograms in rain). They have been selected such that: 1) Shielding and clutter contamination can be neglected, and 2) the highest measurement is considerably below the melting layer (top of 3dB beam at 1.5deg and 24.5km is at 2.5km ASL, while the bright band never fell below 2.7km ASL). Cf. Section 2.4. The numbers of valid data pairs (both values >13dBZ and rain area >50%) used to calculate the six variograms are listed in Table 2.4.

It might seem contradictory on the one hand to postulate intrinsic stationarity (Section 2.3.1), and on the other hand to look for different variograms in neighbouring regions. It is not. First, it is sufficient if the regions for which we determine the variogram are small compared to the overall extent of the analysis. Then, we found the variations of $\mu$ and $\gamma(h)$ within the regions to be negligible. Second, in this section the emphasis is on ranking variograms rather than on their absolute values.

The influence of the increase of the pulse volume with range on variogram estimation is considered to be small compared with the discussed signal. In Section 2.6.2 we find the resulting difference of the variances in the regions Gridone and Vigezzo to be approximately 7%. The estimate is based on data of the MCS five-hour period, which, to a first approximation, represents precipitation of the 96-hour period discussed here. Range influences can be avoided by comparing variograms of regions in the same range interval, e.g. Gridone, Valcolla and Campo dei Fiori.

In the Gridone region reflectivity varied significantly more than elsewhere. The average squared difference between two measurements separated by a given lag is about twice as large as in upstream conditions (region Sesto Calende). This indicates that it has been a preferred region
Figure 2.11: To compare variograms at heights below the melting layer (Fig. 2.13) in upstream and upslope conditions in the Southern European Alps we defined six regions of interest. Geographical coordinates of Monte Lema radar are 8.83deg LON and 46.04deg LAT, corresponding to 707.96km SwissE and 99.76km SwissN, the height is 1625m ASL.
Figure 2.12: Regions for variograms of Fig. 2.14 (thick lines), and region Maggiore (thin line, 914km²). The height interval of the thick-line regions is approximately equal (see Fig. 2.15).
2.7. **Upstream and Upslope Variograms**

![Upstream and Upslope Variograms in Rain](image)

**Figure 2.13:** Variograms in upstream and upslope conditions averaged over 96 hours of September 1999 rain (17-21Sep 25-26Sep 1999). Regions of variograms see Figs. 2.11 and 2.15.

![Upstream and Upslope Variograms up in the Snow](image)

**Figure 2.14:** Upstream and upslope variograms at 4.0-4.4km ASL (above the melting layer). Same period as in Fig. 2.13. Regions of variograms see Figs. 2.15 and 2.12.
### Table 2.4: Statistics of rain variograms (Fig. 2.13).

<table>
<thead>
<tr>
<th>Region</th>
<th>Number of pairs $10^6$</th>
<th>%</th>
<th>$2\gamma(5km)$ dB^2(Z)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sesto Calende</td>
<td>54</td>
<td>8</td>
<td>45.8</td>
</tr>
<tr>
<td>Monte San Primo</td>
<td>45</td>
<td>18</td>
<td>46.6</td>
</tr>
<tr>
<td>Campo dei Fiori</td>
<td>174</td>
<td>13</td>
<td>46.9</td>
</tr>
<tr>
<td>Valcolla</td>
<td>215</td>
<td>16</td>
<td>54.2</td>
</tr>
<tr>
<td>Vigezzo</td>
<td>6</td>
<td>23</td>
<td>58.1</td>
</tr>
<tr>
<td>Gridone</td>
<td>483</td>
<td>37</td>
<td>91.1</td>
</tr>
</tbody>
</table>

### Table 2.5: Statistics of snow variograms (Fig. 2.14).

<table>
<thead>
<tr>
<th>Region</th>
<th>Number of pairs $10^6$</th>
<th>%</th>
<th>$2\gamma(5km)$ dB^2(Z)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sempione</td>
<td>11</td>
<td>15</td>
<td>14.4</td>
</tr>
<tr>
<td>Piora</td>
<td>11</td>
<td>15</td>
<td>17.5</td>
</tr>
<tr>
<td>P. Quadro</td>
<td>9</td>
<td>13</td>
<td>18.7</td>
</tr>
<tr>
<td>Grignone</td>
<td>12</td>
<td>18</td>
<td>19.1</td>
</tr>
<tr>
<td>P. Campo Tencia</td>
<td>12</td>
<td>18</td>
<td>25.2</td>
</tr>
<tr>
<td>Varese</td>
<td>4</td>
<td>9</td>
<td>26.6</td>
</tr>
<tr>
<td>Brianza</td>
<td>7</td>
<td>11</td>
<td>27.1</td>
</tr>
<tr>
<td>Padana</td>
<td>5</td>
<td>6</td>
<td>31.9</td>
</tr>
<tr>
<td>Orta</td>
<td>4</td>
<td>6</td>
<td>35.4</td>
</tr>
<tr>
<td>Càmedo</td>
<td>7</td>
<td>18</td>
<td>45.6</td>
</tr>
</tbody>
</table>
Figure 2.15: Vertical cross section showing the 20 elevation angles of Monte Lema radar (thin lines) and the regions (thick lines) selected for the variograms of Figs. 2.13 and 2.14. The underlain terrain illustrates the need of higher elevation angles to get information above the Alps. The 0.5deg elevation was not used in this direction.

Figure 2.16: Average reflectivity difference in dB(Z) for a separation distance of 5km in different regions, 96 hours of September 1999 rain. Encircled figures are from variograms below the melting-layer, the others from above. Grey-scale for terrain height see Fig. 2.11.
for convective processes. High variation means also low representativity of point measurements and difficulties in runoff modelling. In general, we observe an increase of variation downwind, see last column of Table 2.4.

To extend this analysis towards the crest of the Alps we have to restrict the period to events with high zero-degree isotherms, or, alternatively, to compare variograms calculated with data from above the melting layer (Fig. 2.15). Note that aloft the conditions are different: In the snow variation is generally lower than in rain. But in tall convective cells where large rain drops (and hail) are carried up cores of high reflectivity are surrounded by weak echoes from snow resulting in strong gradients. The convection triggered by the latent cooling of the air during melting (Atlas, 1955; Atlas et al., 1969; Houze, 1997) leading to fallstreaks from the bright band down to the ground, however, is not found in the snow region above.

During the six days the height of the melting layer was too low to have measurements in rain across the Alps, consequently we have to use data from aloft in the snow (Fig. 2.15). The selected regions are shown in Fig. 2.12, the results in Fig. 2.14 and Table 2.5. Maximum variation is again found in the same upslope area, here region Câmèdo. We conclude that in this area upslope triggering of convection plays an important role in precipitation generation. The upstream regions at the edge of the plain of Padana take places 2 to 5 (upstream triggering of convection). The second column of Table 2.5, which shows the number of pairs in percent as a fraction of the total possible number, indicates that upstream rainfall was of short duration. It is interesting to note that the regions close to the main divide (Sempione and Piora) are characterised by weak reflectivity variation.

For comparison, Fig. 2.16 shows the square root of the values at 5km lag of all variograms (below and above the melting layer). These values can be interpreted as the average difference in dB(Z) between two points separated by 5km. At a distance of 5km we expect large gradients in convection. Further, 5km is small enough to obtain an accurate estimate of $2\gamma(h)$ for all regions. The same lag has also been selected for the Tables 2.4 and 2.5.

Figure 2.17 summarises the results of this section comparing variation of reflectivity (2nd moment) with gauge totals (1st moment) along a section from SSE to NNW. Simplified for the 96 hours of heavy rain we can say: Maximum variation approximately coincides with maximum precipitation amount (or with steepest gradient in precipitation?). To the south variation is moderate and gauge totals are small (isolated upstream convection), while to the north variation is weak and gauge totals are moderate to high (persistent stratiform rainfall). Similar results are presented by Harris et al. (1996) who studied multifractal characteristics of orographic precipitation at different locations (gauges) along a transect from the coast to the main divide of the Southern Alps of New Zealand.

A comparison with the partition of precipitation in stratiform and convective as output of meso-scale NWP-models is planned. We want to know how correct does the model predict
2.8. Nugget Variance

It is important to realise that the nugget variance of a regionalised variable $Z(p)$ has nothing to do with $\text{Var}\{Z(p)\}$. The variance of $Z(p)$ is the value to which $\gamma(h)$ tends as $|h| \to \infty$ (see triangle in Fig. 2.8). Cf. Section 2.3. The nugget variance, on the other hand, is the discontinuity at the origin of the semi-variogram, i.e. $\gamma(h)$ as $|h| \to 0$ (Fig. 2.18). It is half the average squared difference of two measurements separated by an infinitesimal small distance, and thus represents the stochastic uncertainty of a single measurement. Only the stochastic part of the uncertainty contributes to the nugget, any systematic difference between the measured and the true reflectivity has no influence.

If we linearly extrapolate the variogram at small lags to zero-lag we get an estimate of the sum of the measurement error and small-scale variability. With small-scale variability we intend the variation at lags smaller than the smallest separation distance in the selected
Figure 2.18: To estimate the nugget variance we plot the variograms of Fig. 2.7 at small lags on a linear abscissa and extrapolate the curves to zero-lag. The nugget variance is half the extrapolated value. Events see Section 2.5 and Fig. 2.7.

region of samples of the size of the pulse volume. For more details see the discussion of the scale of variation in Cressie (1993). Examining 12 variograms of logarithmic reflectivity of 9 precipitation events to the north and to the south of the Alps we found nugget variances ranging from 0.3 to 1.9dB^2(Z). The average value is 1.0dB^2(Z), providing a rough estimate of the average stochastic uncertainty of single measurements. Here, single measurements are polar pixels of 1km x 1deg x 1deg at ranges between 9 and 32km, the 3dB beamwidth is 1deg. Further averaging typically done for the generation of cartesian products reduces this value.

The inphase and the quadrature phase components of the vectorial sum of the complex signals of many drops moving in a random way in the pulse volume are normally distributed (central limit theorem). The standard deviation of logarithmic power of signals with normally distributed inphase and quadrature phase components is $10 \cdot \frac{5}{\sqrt{6}} \log e = 5.6$dB (Marshall and Hittschfeld, 1953). If these fluctuations are the dominant factor contributing to the nugget variance, we can use it to estimate the number of independent samples in one pixel. The value of a polar pixel is the average of 396 samples, 33 pulses in azimuth (the centre of the beam scans one beamwidth) times 1283m-gates in range (post-detection averaging). However, these samples are not independent. The square of 5.6dB divided by 1.0dB^2 from above gives an average number of independent samples in one polar pixel of about 30. For a nugget variance varying between 0.3 and 1.9dB^2(Z), we get 103 (convective) to 16 (stratiform) independent samples, a reasonable result.
2.9 Conclusions

The variograms of Fig. 2.18 let us speculate on a correlation between the degree of variation and the nugget variance, convective precipitation having smaller nuggets. A reasonable explanation could be wind shear and turbulence, which are typically stronger in regions with convection. As a result the decorrelation time is shorter, and the average of successive samples in azimuth and range fluctuates less than in situations with a long decorrelation time. A more detailed analysis is needed to confirm or reject this speculation.

Spatial variation of Alpine precipitation has been quantified by means of variogram analysis using high-resolution reflectivity data. In the Sections 2.3 and 2.4 we presented the background of this geostatistical technique and discussed the specific problems arising when dealing with radar data. A set of typical variograms of Alpine precipitation has been given in Section 2.5. In a nutshell, the answers to the questions listed in the introduction and discussed in the Sections 2.5-2.8 are:

1. The range of variation observed in the various types of Alpine precipitation is large. At a separation distance of 10km the expected difference of reflectivity ranges from 4dB(Z) in stratiform to about 13dB(Z) in convective rain (Fig. 2.8).

During the heavy rain in September 1999 variation in terms of logarithmic reflectivity was weaker above the melting layer than below (Fig. 2.16 and Tables 2.4 and 2.5). Above the melting layer 2\(\gamma\) varied from 14 to 46dB\(^2\)(Z), below in the rain from 46 to 91dB\(^2\)(Z). Examining the variance in terms of logarithmic reflectivity is equivalent to look at the coefficient of variation of linear reflectivity (ratio of the standard deviation to the mean). Thus, even by correcting measurements from aloft by the factor mean(below)/mean(aloft) we can not fully explain the variation observed at the ground. That is to say, measurements from aloft in the snow only in part reflect what happens at ground level. This is an intrinsic limitation of linear profile-correction schemes.

2. Averaging is a common method to reduce the variance. We used variograms to quantify the reduction of variance when averaging measurements in space, integrating radar images in time, or increasing the pulse volume.

Averaging 40min-average stratiform rain over 16, 64 and 256km\(^2\) reduces the variance by 8, 13 and 26\% respectively (Tables 2.1 and 2.3). In precipitation represented by the variogram of the MCS five-hour period, the relative reduction is larger and amounts to 28, 44 and 69\% respectively.
During a five-hour period of a meso-scale convective system the square root of $2\gamma(10\text{km})$ decreases from $7.6\text{dB}(Z)$ using raw data to $2.9\text{dB}(Z)$ for 60min average reflectivity (Fig. 2.7). This corresponds to a factor decreasing from $5.8$ to $1.9$.

Because of the increasing pulse volume, the variance of radar measurements decreases with increasing range. To estimate the order of magnitude we used Krige's formula together with a variogram of linear reflectivity. In precipitation represented by the variogram of the MCS five-hour period the variances of measurements at 9.5 and 31.5km, e.g., would roughly differ by $6\%$. Since the variograms of Fig. 2.7 are based on data between ranges of 9.5 and 31.5km, we conclude that the influence of the pulse volume on the derived variograms can be neglected compared to the observed differences between the various precipitation types.

3. As a consequence of 1. and 2. the representativity of a point observation strongly depends on the type of precipitation (on the time and the location), on the integration time, as well as on the size of the basin. For the 576km$^2$ basin and a 40min period the fractional error, i.e. the expected difference between a point observation and the average basin rate normalised by the overall mean, varies from $11\%$ (variogram of stratiform autumn rain) to $65\%$ (variogram of MCS). See Tables 2.1 and 2.2. In other words, the lack of representativity of a gauge measurement for the average rainfall in a drainage basin ranges from negligible to serious. Analogously we can calculate the fractional error for any variogram and basin.

4. The variogram can be used to estimate the meteorological uncertainty for a given separation distance (Section 2.6.3). Say, for raw data and 60min average reflectivity the square root of $2\gamma(10\text{km})$ is 7.6 and 2.9dB(Z) respectively (see item 2.). Hence, when comparing hourly reflectivity profiles of this type of precipitation of two vertically pointing radars 10km apart one would still have to expect differences of about 3dB(Z). Given the 7.6dB(Z) for raw data, comparing instantaneous profiles would be foolish.

5. We used variograms to look for preferred regions for convection during 96 hours of heavy rain in the Southern Alps in September 1999 (MAP special observing period). We found that close to the crest of the Alps variation was weak (persistent stratiform rain) while maximum variation (frequent convection) occurred in upslope regions (Figs. 2.13 and 2.14).

6. Finally, we determined the nugget variance of 12 variograms (examples see Fig. 2.18). We found values between 0.3 (convective) and 1.9dB$^2(Z)$ (stratiform), giving a rough estimate of the stochastic uncertainty of single radar pixels.

The high-resolution information of radar images is unique for analyses of the spatial continuity of precipitation. The results show that, even under the difficult conditions in a mountainous
region, radar data gives quantitative answers to practical questions related to the spatial continuity. Next steps will be the integration of more events, and the extension to a larger domain incorporating other radars and gauge data.
Chapter 3

Meso-beta Profiles to Extrapolate Radar Precipitation Measurements above the Alps to the Ground-level

3.1 Abstract

In the Alps, the volume visible by a radar is reduced because of ground clutter, elevated horizon and earth-curvature. This often inhibits a direct view on precipitation close to the ground. When using radar measurements from aloft to estimate precipitation rates at ground-level, we must correct for the vertical change of the radar echo, caused by the growth and transformation of precipitation (profile). In this paper we present a robust profile-correction scheme for operational use in complex orography. We aim at correcting for the large errors related to the profile in an Alpine environment: Frequent underestimation caused by the vertical decrease of the radar echo, and occasional overestimation in the bright-band. The profile is determined from volumetric radar data integrated over a few hours within a 70km range of the radar (meso-beta scale).

The correction scheme is verified by comparing radar estimates to gauge measurements of 247 hours of summer and winter precipitation. During the selected period ten gauges collected a total of 3966mm of water. Four concepts to estimate ground-level precipitation are compared: The vertical maximum, the lowest visible echo, estimates corrected with the average event profile, and estimates corrected using the meso-beta profile. Comparisons with the ground truth show that in summer profile-correction considerably reduces the bias and scatter. The root mean square error diminishes by a factor of two. In winter, the improvement is less obvious, but the correction goes in the right direction.

The algorithm is currently being implemented in the operational radar network of MeteoSwiss. Long-term verification is needed after a few years of operation.
3.2 Operational Profile-Correction in the Alps

Measuring precipitation in the Alps is a challenge. Gauge networks have to deal with wind-induced errors, lack of representativity in highly variable precipitation, heating needed to melt the snow, snow burrying the instrument, difficulties in data transfer, unreliable power supply and difficult access for maintenance. The accuracy of quantitative estimates of precipitation by radar, on the other hand, is limited by several factors such as ground clutter, beam occultation and overshooting, profile variation, partial shielding, attenuation, both caused by intervening rain (Delrieu et al., 1991) and a wet radome (Germann, 1999a), as well as Z-R variation (e.g., Joss and Germann, 2000). Current Alpine precipitation maps neither attain the desired precision nor even what would be possible with the actual state of the art. In the operational C-band Doppler radar network of MeteoSwiss calibration and elimination of ground clutter are satisfactorily dealt with (Joss et al., 1998). At present, the dominant source of error is reduced visibility combined with the vertical change of the radar echo (profile), caused by the growth and transformation of precipitation. With reduced visibility we mean the fact that the radar can often not see precipitation close to the ground. This may have several reasons: Ground clutter, beam occultation and overshooting caused by elevated horizon and earth-curvature. Radar sites on mountain tops (Fig. 3.1) improve the horizon, but increase the amount of clutter and reduce the ability to measure close to the ground at short ranges. Locations and heights of the three MeteoSwiss radars (Monte Lema, La Dôle and Albis) are shown in Fig. 3.6.

Figure 3.2 illustrates three types of error related to the profile: Reduced visibility (error A), bright-band contamination (error B), and beam-smoothing (error C). In large parts of the Alps we have to use measurements from aloft to estimate precipitation rates at the ground.

![Scan geometry Monte Lema](image)

**Figure 3.1:** Illustration of the scan geometry and the visibility of Monte Lema radar in the direction of Dissimo (azimuth 297deg).
3.2. Operational Profile-Correction in the Alps

Figure 3.2: Illustration of three types of error related to the profile: Error A: Measurements from aloft used in regions where the lower part of the precipitation is not visible differ from the ground-level rate \( R \). Error B: Measurements in the melting-layer often differ from the ground-level rate \( R \) (bright-band). Error C: Averages over the pulse volume differ from the ground-level rate \( R \), even in perfectly visible regions. Of course error C also occurs aloft. The antenna gain function is indicated in the leftmost plot.

(error A). Therefore, we have to correct for the vertical profile of precipitation (Joss and Waldvogel, 1990). Error B refers to the overestimation caused by the melting layer. The third type of error (error C) would even occur in perfectly visible regions: As the radar beam is not infinitely small, each measurement is an average over a given interval of height. Unless the precipitation rate is constant with height, the average over the pulse volume differs from the value close to the ground.

Within the Swiss Alps the height of the lowest measurement lies between 2 and 5km above sea level (Fig. 3.1). Thus, error A is important almost at all ranges, while error B may be dominant close to the radar only. When ignoring the influence of the profile, the resulting errors are often serious, typically more than 3dB (factor of two) in terms of precipitation rates. If the profile were constant, the correction for error A, B and C would be straightforward. But the variability of the profile in time and space makes profile-correction a difficult task.

A first requirement for radar measurements in an Alpine context is a scan strategy providing data at several heights, such that if a pixel is cluttered or behind a mountain we can use the neighbouring pixels. High resolution in the elevation is also needed to determine vertical profiles. The MeteoSwiss radars make a complete volume scan with 20 elevation angles measuring up to ranges of 230km within 5min. Up to an angle of 9.5deg the elevation increments correspond to the 3dB beam-width (i.e. 1deg). More details see Joss et al. (1998). To obtain a better vertical resolution for a given wavelength and antenna size, Sánchez-Diezma
et al. (2000) propose increments smaller than the 3dB beam-width. This may be a useful addition.

3.2.1 This paper

Up to now too little attention has been paid to the particular situation of profile-correction in complex orography. In this paper we present a robust profile-correction scheme designed for real-time use in the operational radar network of MeteoSwiss in an Alpine environment. With the proposed technique, which is based on profiles on a meso-beta scale (a few hours × 140km), we aim to correct for the large errors occurring in the Alpine context. We correct for reduced visibility (error A) and bright-band contamination (error B) considering variations of the profile at a meso-beta scale. In addition to correcting for errors A and B, the proposed method reduces stochastic errors, since profile-corrections allow to combine measurements in the vertical. With stochastic errors we mean, for instance, variations at time scales smaller than the time between two radar measurements at the same position, or remaining signal fluctuations. This profile-correction scheme is a refinement of the procedure described in Joss and Lee (1995).

Note that the average bias of a radar is not modified by profile-corrections, and must be compensated with long-term adjustment using ground truth (gauges). Such a bias occurs when the radar constant, implemented in the software, differs from the true one, e.g. because of ageing of the hardware or poor estimates of the antenna gain. Another important error source not related to the profile is partial shielding. It is the main reason for the failure of the 'using the lowest echo for precipitation estimates'-approach. Above Dissimo after the obstacle (Fig. 3.1) the power of the third-lowest elevation is approximately half the transmitted power. As a result, precipitation rates are significantly underestimated. The sharp signal diminution at low levels in Fig. 3.3 is in part attributed to partial shielding. In the Alps partial shielding exhibits a complex four-dimensional picture and its correction is difficult. Currently, we use a static polar visibility map for correcting. The concept is described in Joss and Lee (1995).

3.2.2 The Steps of Profile-Correction

The profile-correction consists of the following steps:

1. Determine a representative profile.

2. Calculate on the basis of the profile for each pixel of the 3D radar volume the correction factor needed to extrapolate the measurement down to the ground-level. For poorly visible pixels the correction factor is limited to a maximum value.
Figure 3.3: The lowest 'visible' measurement often underestimates the ground level precipitation rate because of partial shielding. The profile in the figure is the average precipitation rate as a function of height integrating all pixels (also completely shielded ones) in a 28km × 28km region to the north-west of Monte Lema (Fig. 3.4). Time aggregation is controlled by the pwe-function (Section 3.4.2). The time constant is approximately one hour. The decrease of the profile between point P and point Q (−4.1dBR) is caused by shielding (−2.9dBR) and partial shielding (−1.2dBR). 1.2dB in R (rain rate) corresponds to 1.8dB in Z (reflectivity). The figure −2.9dBR is estimated as follows: We assume the precipitation area to be constant between P and Q. Then, the ratio between the observed precipitation areas at the two levels estimates the decrease caused by shielding. The figure −1.2dBR is obtained by subtracting −2.9dBR from −4.1dBR.
3. Apply the correction factor to all pixels. We propose to use all available data in the vertical, rather than considering the lowest visible beam only.

4. Integrate the extrapolated values into the precipitation map and use weighted averaging where for one ground pixel more than one estimate is available. The weights reflect the quality and are calculated considering

   (a) the visibility (pixels that are partly shielded or frequently contaminated by clutter have small weights),
   (b) the distance from the ground (pixels close to the ground have large weights, Joss and Lee, 1995),
   (c) and the position with respect to the melting layer (measurements in the zero-degree layer have small weights).

If the main difference between a local profile and the corresponding meso-beta profile is the intensity of the bright-band, point 4c above reduces the error that results when applying the meso-beta instead of the unknown local profile. The height of the melting layer can be derived from a temperature profile, or by examining the first two moments of the pdf (probability density function) of precipitation rate. In pronounced bright-bands the first two moments are large compared to values below and above (Fig. 3.5).

Since the technique is developed for operational use, it must be suitable for all weather situations. We can not limit ourselves to a certain type of precipitation, for instance, considering widespread stratiform rain only.

It is well-known that the accuracy of precipitation measurements by radar decreases with increasing range (Fabry et al., 1992; Kitchen and Jackson, 1993). Hence, the focus of quantitative estimation of precipitation is at short to medium ranges (up to 120km). Nevertheless, as the correction goes in the right direction, it is continuously applied up to the maximum range of 230km.

A next step will be to consider the influence of the orography (systematic change of profile in the Alps) and, where feasible, profile variation at scales smaller than meso-beta. To do so we adapt the meso-beta profile to the position within the Alps and to the type of precipitation by combining past experience and the 4D-reflectivity information aloft.

### 3.3 What is the Optimum Scale for Profile-Correction?

The optimum scale for profile-correction depends on the meteorological variation of the profile, as well as on the ability to estimate the profile in shielded regions. A decisive point is to select
the right scale: small enough to encounter a significant part of the profile variation, and large enough so that we are able to estimate a representative profile for shielded regions. The question of representativity is closely related to the variability in time and space. The quantity to measure the representativity depends on the application, here the vertical extrapolation of precipitation measurements: We define the estimated profile to be representative, if the error caused by the difference between the estimated and the true profile is small compared to the gain achieved by profile-correction. The gain is measured in terms of improvement of ground-level precipitation estimates.

In a space-time-scale representation, meteorological phenomena are crowded close to a line, each spatial scale being associated with a characteristic time scale (Orlanski, 1975). Thus, from a meteorological point of view, it makes no sense to aggregate data on a scale that is represented by a point far away from that line. Take, for instance, a profile calculated from data of a 5min \( \times \) 100km space-time frame. A 5min time-step would be small enough to resolve the stages of a convective cell. But, as the spatial scale is 100km, the profile is probably a mix of several cells of different stages and of stratiform rain inbetween. Thus, we can just as well select a 120min \( \times \) 100km scale, which is meteorologically equal, but in a statistical sense more robust, because there is a better chance of getting a representative sample when using a 24 times longer period. This is illustrated by the following experiment.

To compare variation in space with variation in time we calculate meso-gamma profiles of 5 summer events including both stratiform and convective precipitation. The terms stratiform and convective refer to precipitation-generating processes reflected by the three-dimensional structure of radar echoes (Houze, 1997). Each event is divided into 8 periods of 60min. Six regions of interest of 600km\(^2\) are chosen (concentric sectors, Fig. 3.4). Thus we obtain \( 5 \times 8 \times 6 = 240 \) profiles (each containing data of 60min \( \times \) 600km\(^2\)). Examples of meso-gamma profiles can be found in Fig. 3.5. We group the profiles using various criteria.

Here, we present the results for two criteria: the width of the pdf below the melting layer (arrangement A) and time (arrangement B). For each arrangement we compare the profile variance within the groups with the total variance of all 240 profiles (analysis of variance, see e.g. Webster and Oliver, 1990). For instance, for the five summer events we find that 65% of the variance of bright-band magnitude (peak divided by ground-level intensity) can be explained by the width of the pdf below the melting layer (arrangement A). The width is a robust parameter for the precipitation type, convective precipitation having larger values (Fig. 3.5). The bright-band magnitude is one of the obvious attributes to describe the type of the profile. But, to examine the variation of the profile, we could just as well use another attribute, such as the ratio of reflectivity below the melting layer and 2km above. As we were saying, arrangement A explains 65% of the variance of bright-band magnitude. If, on the other hand, we divide each event into 2 periods of 4 hours (arrangement B), obtaining \( 5 \times 2 = 10 \) groups of each \( 4 \times 6 = 24 \) profiles, only 41% of the variance of bright-band
Figure 3.4: Regions for meso-gamma profiles: Six concentric sectors of 40km × 36deg to analyse the variability of the profile in space, and two square regions of 28km × 28km for the profiles of the Figs. 3.3 (square to the north-west) and 3.5 (square to the south-east) respectively. Grey-scale for terrain height see Fig. 3.6.

magnitude is explained. Similar experiments have been made with measurements from above the melting layer. We conclude that during summer at the selected scale (60min × 40km × 36deg), profile variation in space (convective cells versus stratiform rain) may be of the same order of magnitude as variation in time (comparing profiles of two 4-hour periods). Thus, only increasing the resolution in time or space does not yield a better resolution of profile variation, but just reduces the representativity of the derived profile.

Table 3.1 gives an overview of the scales of the techniques presented in the literature so far. The 5mm × 140km scale selected by Joss and Lee (1995) was motivated by the simplicity of implementation, since such a profile can be calculated separately for each volume scan.

3.3.1 Why Meso-beta?

For the following reasons we decided to work on a meso-beta scale (a few hours × 140km diameter): a) With meso-beta profiles we can grasp an important part of profile variation, as demonstrated in Section 3.5. Joss and Pittini (1991a) have shown that already seasonal profiles explain an important part. b) In the Alps too little information is available to operationally
3.3. What is the Optimum Scale for Profile-Correction?

Figure 3.5: Meso-gamma profile of convective (left, 2 August 1998, 1000 UTC) and stratiform (right, 3 August 1998, 0720 UTC) precipitation in a 28km x 28km region over Lugano, south-east of Monte Lema (Fig. 3.4). Time aggregation is controlled by the pwe-function (Section 3.4.2). The time constant is approximately 3h and 1h for the convective and the stratiform profile respectively. For each height interval of 500m the average rate (profile) and the probability density function (pdf) have been calculated. The abscissa is logarithmic. The pdf having 15 classes is expressed as a percentage of the total precipitation area, which is defined as the area covered by precipitation exceeding 0.16mm/h (Z>13dBZ). On the ordinate of the pdf 0% and 10% is indicated. The figure in parenthesis indicates the precipitation area: 36%, e.g., tells that within the corresponding height interval 36% of all pixels have R>0.16mm/h. Representing the pdf as a function of height is similar to the CFAD of Yuter and Houze (1995).
Table 3.1: Scales of profile-correction schemes.

<table>
<thead>
<tr>
<th>Method</th>
<th>Aim</th>
<th>Profile Scale</th>
<th>Single\textsuperscript{a}</th>
</tr>
</thead>
<tbody>
<tr>
<td>Smith (1986)</td>
<td>Bright-band correction</td>
<td>15min × 15deg</td>
<td>no</td>
</tr>
<tr>
<td>Joss and Pittini (1991b)</td>
<td>Profile-correction</td>
<td>10min × 140km × 140km</td>
<td>yes</td>
</tr>
<tr>
<td>Koistinen (1991)</td>
<td>Profile-correction</td>
<td>24h × 90km diameter</td>
<td>yes</td>
</tr>
<tr>
<td>Kitchen et al. (1994)</td>
<td>Profile-corr. (widespread rain)</td>
<td>10min × 5km × 5km</td>
<td>no</td>
</tr>
<tr>
<td>Divjak (1995)</td>
<td>Profile-correction</td>
<td>60min × 120km diameter</td>
<td>yes</td>
</tr>
<tr>
<td>Andrieu and Creutin (1995) and Andrieu et al. (1995)</td>
<td>Profile-correction</td>
<td>60min × 200km diameter</td>
<td>yes</td>
</tr>
<tr>
<td>Joss and Lee (1995)</td>
<td>Profile-correction</td>
<td>5min × 140km × 140km</td>
<td>yes</td>
</tr>
<tr>
<td>Smyth and Illingworth (1998b)</td>
<td>Profile-correction</td>
<td>radar pixel</td>
<td>no</td>
</tr>
<tr>
<td>Vignal et al. (1999)</td>
<td>Profile-correction</td>
<td>60min × 10–30km × 15deg</td>
<td>no</td>
</tr>
<tr>
<td>Seo et al. (2000)</td>
<td>Profile-corr. (widespread rain)</td>
<td>not specified</td>
<td>yes</td>
</tr>
<tr>
<td>This paper</td>
<td>Profile-correction (Alps)</td>
<td>a few h × 140km diameter\textsuperscript{b}</td>
<td>yes</td>
</tr>
</tbody>
</table>

\textsuperscript{a}Single profile for whole radar range.
\textsuperscript{b}Visible part only.

estimate a representative profile in shielded regions at scales smaller than meso-beta. Operationally means non-stop all year round and in the whole radar volume. c) Aggregation in space and time is needed to determine vertical profiles using data of a volume-scanning radar. d) Both, spatial and temporal averaging reduce stochastic errors and micro-scale signatures, and preclude problems caused by slanted precipitation, described by Fabry and Zawadzki (1995).

3.4 Method to Calculate Meso-beta Profiles

The profile is defined as the average precipitation rate versus height within a given space-time frame. The operational algorithm works with a resolution of 200m in the vertical, and a time step of 5min. To reduce the amount of data when analysing long events in post-processing (e.g. for this paper), we use 500m and 10min respectively. Since cartesian radar products would introduce resampling errors, profiles are derived from polar data only. Precipitation rates are derived from reflectivity by means of a constant Z-R relation \((Z = 316 R^{1.5})\). This Z-R relation was derived for rain. Therefore, the profile in the snow reflects an equivalent rain rate rather than the snow rate. This is of little importance if precipitation at the ground is liquid. Then, it is sufficient if the same Z-R relation is used for determining the profile and for the measurements extrapolated to the ground. In case of snowfall (in the mountains or during winter also in the lowlands), we may implement a correction to account for the different Z-R relation of snowfall. However, an attempt in Finland of real-time adjustment of
3.4. Method to Calculate Meso-beta Profiles

the Z-R relation according to the water-phase of the hydrometeors did not lead to a significant improvement (Saltikoff et al., 2000). Other errors were dominant.

3.4.1 Aggregation in Space

To determine meso-beta profiles, precipitation is averaged over space and time. Each polar pixel within a given region is weighted by the area of the corresponding pulse volume. Thus, we obtain an area-weighted average of the precipitation rate. Meso-beta profiles are calculated from data of a 70km-radius cylinder centred over the radar station (Fig. 3.6). We only consider the region that is well-visible and clutter-free.

3.4.2 Aggregation in Time: pwe-function

On the time-scale we use a precipitation-volume-weighted exponentially-decaying function, hereinafter referred to as pwe-function. The weight of a volume-scan continuously decreases with time (Fig. 3.7). The decay 'constant' \( \alpha_t \) is variable and depends on the amount of new information available. It is maximum (fast adapting to the new profile) when there is precipitation in the whole region (all pixels have \( Z > 13 \text{dBZ} \)), and zero (profile is kept constant) in the absence of precipitation. The pwe-function is

\[
\bar{p}_t = (1 - \alpha_t)\bar{p}_{t-1} + \alpha_t p_t
\]

where \( \bar{p}_t \) is the average and \( p_t \) the instantaneous profile at time step \( t \); the decay constant \( \alpha_t \) is obtained from

\[
\alpha_t = \frac{c_t}{C_0 + c_t}
\]

where \( c_t \) is the precipitation volume at time step \( t \). The precipitation volume is defined by means of a lower reflectivity threshold of 13dBZ (corresponding to 0.16mm/h). \( C_0 \) is a constant and determines the minimum time constant of the exponential decay. If \( C_0 \) is set to zero, there is no integration in time (time constant \( \tau_t = 0 \)) and simply \( \bar{p}_t = p_t \). A large value, on the other hand, corresponds to a large time constant. The time constant \( \tau_t \) is defined as the time when an initial weight of a profile has decreased by a factor of \( 1/e \)

\[
\tau_t = -\Delta t[\ln(1 - \alpha_t)]^{-1}
\]

where \( \Delta t \) is the time step. Physically \( C_0 \) is a volume. We propose to set

\[
C_0 = 2C_v
\]

where \( C_v \) is the visible part of the region of spatial aggregation (here the 70km-radius cylinder). Table 3.2 shows the time constant of the pwe-function given \( C_0 = 2C_v \) and \( \Delta t = 10\text{min} \) for several values of \( \alpha_t/C_v \).
Figure 3.6: Locations of the operational MeteoSwiss radars. The C-band Doppler radars Monte Lema, La Dôle and Albis are located on mountain tops at 1625m, 1675m, and 925m ASL respectively. Meso-beta profiles for operational profile-corrections are calculated from data of a 70km-radius cylinder centred over the radar station. The dash-dotted line in the Monte Lema region separates the well-visible (south) from the shielded (north) subregion. The well-visible subregions of La Dôle and Albis have not been defined yet.
3.4. Method to Calculate Meso-beta Profiles

Table 3.2: Time constant of pwe-function for several $c_t/C_0$. Time-step is 10min, $C_0 = 2C_t$.

<table>
<thead>
<tr>
<th>$c_t/C_0$</th>
<th>1</th>
<th>0.1</th>
<th>0.01</th>
<th>0.001</th>
</tr>
</thead>
<tbody>
<tr>
<td>Decay constant $\alpha_t$</td>
<td>0.33</td>
<td>4.8e-2</td>
<td>5.0e-3</td>
<td>5.0e-4</td>
</tr>
<tr>
<td>Time constant $\tau$</td>
<td>25min</td>
<td>3h</td>
<td>25min</td>
<td>1day</td>
</tr>
</tbody>
</table>

We can rewrite Eq. (3.1) as

$$\bar{p}_t = \sum_{i=1}^{t} \alpha'_{i,0} \bar{p}_i$$

using

$$\alpha'_{i,t} = \alpha_i \prod_{j=i+1}^{t} (1 - \alpha_j)$$

Figure 3.7 illustrates the pwe-function: The three plots (a), (b), and (c) show $c_t$, $\alpha_t$ and $\alpha'_{t,0}$ respectively as a function of time step $t$.

In the profile-correction technique proposed by Seo et al. (2000) time aggregation is controlled by a memory span parameter. For widespread rainfall with a uniform profile they suggest to set the parameter to a small value, in which case the resulting profile is an average over a small period. For intermittent rainfall the memory span parameter is set to a large value in order to smooth the high variability of the profile. The approach is similar to what we propose above (and in Germann and Joss, 1999b) in that aggregation time is variable. Here, we give a rule how to determine this parameter in an operational context.

3.4.3 Arguments for this Technique

Several arguments led to this definition of meso-beta profiles:

1. The pwe-function provides a well-defined profile at any time. At the very beginning of a new precipitation event, for instance, we need not to wait an hour or more to have enough measurements to calculate a well-defined profile. Instead, as a first guess, we take the profile of the most recent precipitation, and, step-by-step, adapt it to the current weather situation by incorporating new measurements. The same is valid for precipitation interrupted by short breaks.

2. Using this technique only small changes can occur between successive profiles. As a consequence, the corrected radar images will exhibit no discontinuities when animated.
Figure 3.7: Illustration of pwe-function. Plot (a) depicts the precipitation volume $c_t$ seen by the radar in the well-visible part of the 70km radius cylinder. Data is from a 36-hour period of Monte Lema ending at 2100 UTC, 3 August 1998. Figure 3.8 shows $c_t$ of the whole event (83 hours). Time step is 10min. The ordinate ranges from $10^3$ (values smaller than $10^3$ correspond to remaining clutter, i.e. no rain) up to the maximum possible value $C_v$. The corresponding decay constant is given in plot (b), while plot (c) shows the weights used to calculate the meso-beta profile at time step $t_0$ (3 August 1998, 2100 UTC). The integral of $\alpha^i_{t,t_0}$ is 1. The weights of the last four hours make up 80%. Weights smaller than $10^{-3}$ do not significantly contribute to the mean but reflect past precipitation. The abscissa is time $t$ indicated in hours before time $t_0$, though, $t$ and $t_0$ are strictly speaking a number of time steps rather than a time (Section 3.4.2).
3. The resulting precipitation estimates are unbiased in the space-time frame in which the profile has been calculated.

4. On a meso-beta scale in the Alps, precipitation growth may predominantly be stratiform, convective, or any combination of both. For long-term precipitation totals both processes are important. The proposed technique is suitable for all types of Alpine precipitation, and requires no stratification (stratiform-convective). It combines the various profiles of the meso-beta space-time frame weighted according to their contribution.

5. The method is stable, simple to implement in an operational context, and has low processing costs. Note that Eq. (3.1) is recursive.

6. In contrast to schemes that have been trained with a limited data set, for instance, those using parametrised profiles, the proposed scheme is self-adaptive. We make no assumption about the shape of the profile, about the importance of different precipitation growth processes, or about the frequency of certain weather types.

7. Using the same technique we can also calculate profiles on meso-gamma scales (Fig. 3.5). We may have to select a smaller region and adjust $C_0$ by changing the factor in Eq. (3.4).

The meso-beta profile defined above is smoothed by the antenna gain pattern, and thus differs from the true profile, that would be seen by a radar with an infinitely small beam-width and pulse-length. Andrieu and Creutin (1995) and Andrieu et al. (1995) propose an inverse method for profile identification. The identified profile is an estimate of the true profile. A similar approach is followed by Seo et al. (2000). Vignal et al. (1999) extend the algorithm of Andrieu and Creutin (1995) to volume scan data, which allows to retrieve local profiles (60min × 400km²) in well-visible regions. These methods are computationally expensive. A comparison of the improvement of rain estimates using no profile, a climatological, a mean, and identified local profiles can be found in Vignal et al. (2000). The mean profile is an hourly average in a 70km-radius cylinder, and is to some extent comparable to the meso-beta profile defined above. The comparison has been performed in the well-visible part of the radar volume of Albis (SW-N-NE). They found the reduction of the fractional standard error (FSE) when correcting for the profile to be significant: an FSE of 25% applying the mean profile compared to 44% without correction. Using identified local profiles further reduces the FSE to 23%.

For operational profile-correction in the Alps we use meso-beta profiles rather than sophisticated profile identification. The reasons are:

8. The advantage in using sophisticated techniques is limited to well-visible regions, and is relatively small (Vignal et al., 2000), while processing costs are high.
In the Alps, the lower part of the atmosphere is not visible. Therefore, we can not measure the full profile where we want to apply it. We have to estimate it. Then the dominant error is often introduced when extrapolating profiles of a well-visible region to a shielded region.

Compared to meso-beta profiles estimates of the true profile better represent the sharp, narrow shape of the bright-band. For profile-correction, however, it is safer to use profiles with smooth bright-bands (Fig. 3.10). This because a narrow bright-band slightly displaced in height may cause serious damage. In the Alps the height of the melting layer varies both in space (air-masses following the terrain) and time (temperature fronts). Hence, to correctly consider the narrow shape of the bright-band, we would need estimates of the profile at a scale of less than 30min × 10km.

### 3.4.4 Limitations

The presented correction scheme has two major limitations: The influence of the orography on the precipitation mechanism is not fully considered, and profile variation at scales smaller than meso-beta are averaged.

(a) Influence of the Orography

The influence of the orography on precipitation is threefold (Houze, 1999):

a) Dynamics — role of wind speed and direction and role of stability and moisture stratification in determining whether air flows easily up and over terrain or is blocked, and in determining the role of convection in the precipitation processes.

b) Microphysics — determining whether warm cloud processes (coalescence) are key in the precipitation growth or whether ice processes are critical. This determines whether water falls out immediately over slopes or is advected over ridges and peaks.

c) Topography — in particular the spatial scale of the terrain features. The entire Alpine barrier modifies the flow within the baroclinic waves and frontal structures that produce the precipitation. The broad indentations such as the Lago Maggiore region provide a mesoscale focus of convergence. Individual river valleys appear from the DOW data to be outlets of return flow during periods of heavy precipitation (Steiner et al., 2000). DOW stands for Doppler on Wheels, one of the research radars set up in the southern Alps during the special observing period of the Mesoscale Alpine Programme (MAP, Bougeault et al., 2000).

As a consequence, the profile systematically varies depending on the position within the Alps, and the profiles measured in the vicinity of Monte Lema are only in part representative for the rest of the radar volume.
Evidence in support of this limitation is given in Germann and Joss (2000b). There, we found parameters describing the type of precipitation to systematically vary when going from the foothills of the Southern Alps towards the main divide. The profile probably does so too. This is consistent with the fact that, after correction for meso-beta profiles, radar measurements still underestimate gauge totals in Alpine valleys. We expect low-level growth to be maximum in upslope regions where the warm and humid low-level flow towards the Alps — in the Southern Alps usually from SE or S — is forced to rise. For the UK radar network Kitchen et al. (1994) estimate an orographic enhancement term based on studies of Hill (1983) combining the low-level wind field with a terrain model. Preliminary results of the Mesoscale Alpine Programme suggest that the situation in the Alps is more complex. To take the orographic influence into account we have not found a method yet that is robust enough for operational use.

(b) Profile Variation at Smaller Scales

Profile variation can be found down to the space-time scale of single instantaneous radar measurements. Such micro-scale profiles exhibit variations caused by drop sorting, convective cells, up- and downdrafts, fallstreaks, and remaining signal fluctuations, etc. They look to some extent random, and their interpretation is difficult, unless detailed in-situ measurements are available. All types of variation at scales smaller than meso-beta are ignored by the profile-correction scheme presented here.

3.4.5 Implementation

Quality checks are needed to prevent contamination with bad information, such as clutter during fine weather, or void radar images in case of short power failures (ups batteries often supply the processors but not the magnetron).

Immediately after a reboot of the radar system, we have the following choices to obtain a profile for correction: a) use a climatological profile, b) restore $\bar{p}$, saved previously, or c) make a first guess with the little information available from the first volume scan after the reboot. When choosing the solution c), $C_0$ of Eq. (3.2) is replaced by $C'_0$ given by

$$C'_0 = \begin{cases} C_0 & \text{if } C_0 < C_m \\ C_m & \text{otherwise} \end{cases}$$

(3.7)

with

$$C_m = (t - 1) \sum_{i=1}^{t-1} \alpha_{i,t-1} c_i$$

(3.8)

Thus, $C'_0$ is calculated according to Eq. (3.4) but has a maximum value of $C_m$ corresponding to the weighted precipitation volume aggregated so far (Eq. (3.8)).
We propose to save together with the profile-corrected radar image also the precipitation volume \( c_t \) and the profile \( p_t \). The two parameters make later quality control easier, and are a basis for climatological studies and the retrieval of further synoptic information such as the height of the zero-degree isotherm.

### 3.5 Verification

To verify the improvement achieved by meso-beta profile-correction, we compare four radar estimates of ground-level precipitation rates with gauge measurements. The radar estimates are: The vertical maximum, the lowest visible, estimates obtained by correcting for the average event profile, and estimates obtained by correcting for meso-beta profiles. For all four radar estimates we only consider pixels that are a) according to the clutter map, at least in 90% of the cases clutter-free, and b) according to optical visibility, at least 80% visible. This is particularly important when using the lowest visible measurement. It is better than just taking the lowest echo, which is often partially shielded or contaminated by ground clutter. While the event profile is an average over the whole event, the meso-beta profile varies in time as described in Section 3.4.2. Thus, the event profile is something between a seasonal profile and a meso-beta profile. The profiles are based on the same data.

The event-profile estimate is identical to the meso-beta-profile estimate, except that instead of the meso-beta profile the average profile of the whole event is used. The two profiles are based on the same data.

Two summer events of 83 hours and 128 hours respectively, and a winter event of 36 hours have been selected. This makes a total of 247 hours of precipitation. With the radar-gauge comparison we aim to answer two questions:

1. Which of the four techniques most effectively reduces the scatter between radar and gauge measurements? That is, we ignore any remaining bias and only look at the scatter.

2. Say, we define four radar products using the four techniques described above. Which product is closest to the observations? Here, we also take into account the remaining bias.

A description of the events is given in Section 3.5.1. In Section 3.5.2 we introduce the statistics used to answer these questions and discuss the results when they are applied to the 247 hours or precipitation.
### Table 3.3: Statistics of 6-hour gauge accumulations of summer and winter precipitation.

<table>
<thead>
<tr>
<th>Season</th>
<th>Period</th>
<th>E( { G } ) mm</th>
<th>Var( { G } ) mm(^2)</th>
<th>Max( { G } ) mm</th>
<th>( \geq 1) mm</th>
<th>( \geq 3.3) mm</th>
<th>( \geq 10) mm</th>
<th>( \geq 33) mm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer(^a)</td>
<td>31Jul98-3Aug98, 17-20/25-26Sep99</td>
<td>211</td>
<td>17.8</td>
<td>287</td>
<td>84.7</td>
<td>207</td>
<td>174</td>
<td>116</td>
</tr>
<tr>
<td>Winter</td>
<td>5-7Nov99</td>
<td>36</td>
<td>8.9</td>
<td>50</td>
<td>27.9</td>
<td>31</td>
<td>23</td>
<td>11</td>
</tr>
</tbody>
</table>

\(^a\)The two summer events are pooled for radar-gauge comparison.

### 3.5.1 247 Hours of Precipitation

The first of the three 'events' is an 83-hour period starting at 1000 UTC on 31 July 1998. The precipitation volume as used in the pwe-function (Section 3.4.2) is depicted for the whole period in Fig. 3.8. Precipitation of the first three days is dominated by convective cells moving from south-west toward the Alps, while daily accumulated PPIs of the fourth day show uniform precipitation and features indicating a pronounced bright-band (stratiform). A PPI (plan position indicator) is a representation on a horizontal surface of radar measurements made at a constant elevation angle. Figure 3.10 shows a time serie of six meso-beta profiles in the early morning of 3 August, calculated as described in Section 3.4. The first profile at 0020 UTC has a weak vertical gradient and no bright-band. It reflects the vertical echo-decrease of the convective precipitation in the evening of the day before. According to Eq. (3.2) the meso-beta profile is held constant in the absence of rain. Then, as in the early morning of 3 August stratiform precipitation enters the 70km-radius region, the meso-beta profile fast adapts to the new situation.

The second event comprises 128 hours of September 1999 rain (MAP IOPs 2a, 2b and 3). The third, which corresponds to MAP IOP 15, is a stratiform event in November 1999 and lasts 36 hours.

Radar estimates are compared with gauge measurements on the basis of 6-hour accumulations. We only take into account intervals during which the gauge has collected at least 1mm. The two summer events are pooled for radar-gauge comparison, whereas the winter event is treated separately. For comparison we select 10 gauges to the north of Monte Lema (Fig. 3.9). During the summer and winter events, the gauges collected a total of 3690mm and 276mm of water respectively. This meso-beta profile-correction technique aims at improving quantitative estimates of precipitation in the Alps by correcting for error A and B (Section 3.2.1). Accordingly, we choose gauges at locations where 1) error A and B are important, and 2) the quality of radar measurements nevertheless allows quantitative estimates of precipitation. Table 3.3 lists
Figure 3.8: Precipitation volume (as in plot (a) of Fig. 3.7) of an 83-hour event in summer 1998. Power failure on 1 August at 1540 and 2200 UTC caused a few void PPIs, see spikes in the figure. The corresponding profiles get small weights only, according to Eq. (3.2).

Figure 3.9: Locations of gauges used for the verification of meso-beta profile-correction. The height of three mountains is also indicated: Pizzo Campo Tencia (3072m), Cima della Laurasca (2195m) and Monte Tamaro (1962m). Monte Lema is at 1625m. Grey-scale for terrain height see Fig. 3.6.
3.5. Verification

Figure 3.10: Time-series of meso-beta profiles, 3 August 1998, 0020 UTC (convective) till 0750 UTC (stratiform). Details see caption of Fig. 3.5.
3.5.2 Radar-Gauge Comparison

To answer the two questions asked at the beginning of this section we calculate for each of the four radar estimates the following statistics:

\[
AF = \frac{E\{G\}}{E\{R\}} \quad (3.9)
\]

\[
RMSE = \left[ E\left\{ (G - R)^2 \right\} \right]^{0.5} \quad (3.10)
\]

\[
ARMSE = \left[ E\left\{ (G - \frac{E\{G\}}{E\{R\}} R)^2 \right\} \right]^{0.5} \quad (3.11)
\]

where \(E\{\}\) is the expectation operator, and \(G\) and \(R\) are the 6-hour accumulations of the 10 gauges and the corresponding radar estimates respectively. \(AF\) stands for adjustment factor, and \(ARMSE\) for the root mean square error of the estimates adjusted using \(AF\). With 'error' we mean the differences between \(R\) (estimate) and \(G\) (reference). The \(ARMSE\) provides an answer to the first question, while answering question two requires calculation of the \(AF\) and the \(RMSE\). Radar estimates are averaged over areas of \(3 \times 3\)km\(^2\) centred over the location of the gauges.

With the currently implemented radar constant the Monte Lema radar underestimates precipitation, even in well-visible regions at short range. To compensate for this, all radar measurements are first multiplied by a factor, which is determined from two gauges at good visibility close to the radar (Fig. 3.9). The factors are 1.49 and 1.84 for 1998 and 1999 respectively.
Table 3.4: Radar-gauge comparison in summer and winter precipitation.

<table>
<thead>
<tr>
<th>Radar estimate</th>
<th>AF</th>
<th>RMSE mm</th>
<th>ARMSE mm</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Summer (211 hours)</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Vertical Maximum</td>
<td>0.81</td>
<td>34.6</td>
<td>27.4</td>
</tr>
<tr>
<td>Lowest Visible</td>
<td>2.92</td>
<td>17.7</td>
<td>24.8</td>
</tr>
<tr>
<td>Event Profile</td>
<td>1.92</td>
<td>14.9</td>
<td>15.0</td>
</tr>
<tr>
<td>Meso-beta Profiles</td>
<td>2.22</td>
<td>14.9</td>
<td>14.2</td>
</tr>
<tr>
<td><strong>Winter (36 hours)</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Vertical Maximum</td>
<td>3.34</td>
<td>9.1</td>
<td>10.8</td>
</tr>
<tr>
<td>Lowest Visible</td>
<td>6.62</td>
<td>10.0</td>
<td>12.7</td>
</tr>
<tr>
<td>Event Profile</td>
<td>3.42</td>
<td>9.3</td>
<td>10.7</td>
</tr>
<tr>
<td>Meso-beta Profiles</td>
<td>1.53</td>
<td>9.2</td>
<td>12.2</td>
</tr>
</tbody>
</table>

We use two factors, because in August 1999, shortly before the summer 1999 event, a new version of the operational clutter-elimination algorithm has been implemented. It significantly reduces the amount of ground echoes at low elevations. The changes in the procedure for clutter elimination explain and justify the larger factor for 1999. Note that applying these factors for pre-adjustment does not affect the ARMSE used to compare the four techniques. The AF and the RMSE, however, are changed. But, as we plan to correct the operational radar constant to compensate for the overall underestimation in the near future, the calculated AF and RMSE are realistic estimates of what will soon be achieved.

Figure 3.11 shows scatterplots of gauge measurements versus the four radar estimates for 6-hour accumulations of the 211 hours of summer precipitation. The AF, the RMSE and the ARMSE, calculated separately for summer and winter precipitation, can be found in Table 3.4.

Let us first look at the summer data. The scatterplot of the vertical maximum shows two branches: one below the diagonal where $R$ underestimates $G$, and one above with overestimations up to a factor of 8 ($G=39\text{mm}, R=308\text{mm}$). The lower branch originates from gauges at severely shielded locations. Radar measurements are from aloft in the snow, and contamination by the bright-band or clutter is unlikely. The overestimation in the upper branch is caused by the bright-band, by remaining clutter, and by the maximum-algorithm. The vertical maximum has both, largest RMSE and ARMSE. The second radar estimate, the lowest visible, may be a solution in the absence of shielding and partial shielding, and in convective cores where the vertical decrease of the echo is weak (points close to the diagonal). But elsewhere, it seriously underestimates ground-level precipitation rates, leading to the points close to the abscissa. The two techniques correcting for the profile significantly reduce both, the overall scatter, and the underestimation in shielded regions, and have an AF smaller than that of the lowest visible. The somewhat smaller AF of the event profile is deceiving. It results from under-correcting
echoes from the bright-band when applying the average event profile with a weak bright-band. Except for the upper branch of the vertical maximum, all radar estimates underestimate ground truth. This is explained by the influence of the orography. Profiles measured in the foothills of the Alps (Fig. 3.6) do not reflect the orographic enhancement of precipitation that occurs below the lowest radar measurement in the Alpine valleys (Section 3.4.4).

For the winter data we can state the following: All four techniques underestimate ground-level precipitation rates. The most serious underestimation occurs when using the lowest visible measurement (a factor of 6.6). The absolute errors (RMSE and ARMSE) are smaller than in summer, but so are the precipitation amounts. Therefore, in winter the fractional errors (RMSE and ARMSE divided by $E\{G\}$ of Table 3.3) are considerably larger, except for the vertical maximum. For the ARMSE of the meso-beta-profile estimate, e.g., we obtain 80% and 137% for summer and winter data respectively. When looking at fractional errors, it is important to take note of the reference. Here, we consider all data pairs with $G \geq 1\text{mm}$. If we increase this threshold, the fractional error will diminish. The 80% from above, for instance, diminishes to 71% and 65% when using a threshold of 5mm and 10mm respectively. It depends on the type of application whether the absolute or the fractional error is crucial. The AF, the RMSE and the ARMSE of the vertical maximum and the event-profile estimates are almost identical. Applying meso-beta profiles results in a smaller AF but a somewhat larger scatter. The overall underestimation is explained by the fact that all winter radar measurements are made above (or at the top) of the melting layer. As a result, the measured profiles applied in the profile-correction do not reflect echo-growth in the melting layer and below.

The scatter between radar estimates and ground truth observed in the Alps is large compared to results of similar studies in not-shielded regions. For nine summer events and gauge data of the well-visible region of Albis, Vignal et al. (2000) found fractional deviations of 44% and 23% using uncorrected and profile-corrected radar estimates respectively. This is not surprising when considering the difficulties that arise in an Alpine environment (Section 3.2). The scatter can be reduced to some extent by integrating radar estimates in time and/or space.

### 3.6 Conclusions

A profile-correction scheme for operational use in complex orography is presented. The goal is to improve quantitative estimates of ground-level precipitation rates in an Alpine environment. Profiles are calculated on a meso-beta scale. In space, we combine radar measurements from a 70km-radius cylinder centred over the radarsite. Integration in time is controlled by the pwe-function (precipitation-volume-weighted exponentially-decaying function, Section 3.4.2). The time constant is of the order of $10^2\text{min}$. The resulting improvement of precipitation estimates reflects that the meso-beta profile explains an important part of profile variation.
The presented method for profile estimation is robust, and suitable for operational use in complex orography. It continuously yields a statistically well-defined profile. Only small changes occur between profiles used to correct successive scans. This avoids discontinuities in the corrected radar images when animated.

We have compared four techniques for the estimation of ground-level precipitation rates in complex orography: The vertical maximum, the lowest visible, correcting for the average event profile, and correcting for meso-beta profiles. For verification we use measurements of ten gauges in the Alps (Fig. 3.9) during 247 hours of summer and winter precipitation. In summer, profile-corrections cause a distinct improvement compared to the two conventional techniques, vertical-maximum and lowest-visible. The root mean square error diminishes by a factor of two (Table 3.4 and Fig. 3.11). Contamination by the bright-band and underestimation in shielded regions are both reduced. Correcting using the meso-beta profile gives the best result. In winter, the improvement achieved with profile-correction is less obvious, because we lack the lower part of the profile (Section 3.5.2).

At this stage, profile-corrected radar images in the Alps provide a better estimate of the precipitation pattern, but still underestimate the amounts (by a factor of 2 in summer, and 1.5 in winter). This remaining bias is attributed to systematic differences between the profile measured in the foothills and the profile in the Alps (summer), and to the ignorance about the low-level growth (below the lowest measurements, especially in winter). Future work and experience after a few months of operation in the radar network of MeteoSwiss will show to what extent this can be considered. In winter, when the precipitation is shallow, the radar sometimes completely misses the precipitation. Then, the best profile-correction scheme can not help.
Chapter 4

Radome Attenuation — A Serious Limiting Factor for Quantitative Radar Measurements?

4.1 Summary

A cover of water, wet snow and/or ice on the radome, the protection sphere around the radar antenna, may cause serious signal attenuation. This is confirmed both by theoretical models and experimental measurements. An illustrative example is the often observed lighthouse effect when a storm passes the radar station. For qualitative work, such as detecting and tracking precipitation areas, the reduction of the signal may be acceptable, whereas for quantitative precipitation estimation it may have disastrous consequences. Modeling the complicated structure of the water cover on the radome is a rather difficult task. Thus theoretical values of radome attenuation have to be compared with measurements.

This paper estimates radome attenuation under natural conditions using radar measurements of precipitation. Seven precipitation systems observed with the Swiss C-band Doppler radars have been analysed. Reflectivity measurements before, during and after rain on the radome are compared. The average signal loss within the precipitation area gives a quantitative estimate of the phenomenon. Complementary information is obtained from investigating reflectivity probability distributions and reflectivity transformation matrices. Radome attenuation ranges up to 5.4dBZ two-way in moderate rain, and thus reduces the precipitation estimates to less than half. Results of theoretical models lead to the same order of magnitude.

It has not yet been shown to what extent errors can be reduced. Since radome attenuation shows a complex spatial and temporal variation its correction will be difficult. We propose to carefully interpret data measured while the radome may be covered by water and to further investigate the phenomenon.
4.2 Introduction

For qualitative work such as detecting and tracking precipitation the today's meteorological radar gives reliable information. Quantitative interpretation of radar data, however, is far more demanding because of the precision required (usually a fraction of 1dB two-way). In order to use data of the Swiss weather radar network for quantitative precipitation estimation large efforts have been made. Steps towards this aim include calibration, clutter elimination, profile and beam shielding correction, and adjustment with rain gauges (Joss, 1996b). Attenuation during propagation (mainly caused by intervening rain) and its correction has been the subject of many studies (see e.g. Hitschfeld and Bordan, 1954; Geotis, 1975; Delrieu et al., 1991). This paper treats signal attenuation caused by water on the radome.

The radome is a spherical cover that protects the antenna from mechanical stress by precipitation, ice and wind. By using the right material the radome itself does not cause serious attenuation (below 0.5dB one-way). But if the radome is covered by a film of water or — which is even worse — by a mixture of water and ice, the signal is attenuated by many times the dry radome loss. The effect depends on the transmission frequency: it is certainly serious at shorter wavelength (X-band, 3cm) and usually negligible at long wavelength (S-band, 10cm).

To determine the signal loss caused by a wetted radome of satellite communications antennas various attempts have been made (see for example Cohen and Smolski, 1966; Ruze, 1965, 1966; Anderson, 1975). However, those systems use higher transmission frequencies and are probably less precision-demanding than precipitation radars. Austin (1987) discusses physical factors that influence the relation between the measured radar reflectivity and surface rainfall. During one of the twenty analysed storms she found a drastic decline of the radar measurements. It was the only storm the data of which were taken with a C-band radar and the decline has been explained by an attenuating water cover on the radome.

The main goal of this study is to estimate radome attenuation under natural conditions. Section 4.3 discusses theoretical aspects of radome attenuation. In section 4.4 first we describe the proposed technique. Then we present the results of the studied precipitation events observed with the Swiss C-band radars (transmission frequency is 5.4GHz). Finally in section 4.5 the conclusions are listed and a rough comparison of the results with modeled values is made.

4.3 The Phenomenon

An electromagnetic wave that hits a water-layer covering the radome is partially reflected (about 60% of the power, Joss, 1996a) and shifted in phase because of the loss tangent (the
ratio between the imaginary and the real part of the complex dielectric constant). The rest enters the water-layer and is reflected back and forth within the layer many times (multiple reflection). As a result the attenuation is much larger and more complicated than calculated without considering multiple scattering. An irregularly thick water film causes variable attenuation and phase-shifting and therefore de-focusing of the picture. After Joss (1996a) at a transmission frequency of 5.4GHz a water film with a thickness of 0.1mm [0.3mm] leads to one-way attenuation of 2dB [6dB] and a phase shift of $-0.05\lambda$ [$-0.1\lambda$]. In the case of a mixture of water and ice the reduction of the signal is even larger. Ice is a nearly lossless dielectric. Together with water the mixture is as lossy as water but can be much thicker than the water film. The result for quantitative precipitation estimation may be disastrous. Anderson (1975), e.g., measured rapidly increasing losses while the snow or ice on the radome was melting.

To calculate attenuation of a water film we must know the thickness of the film as a function of the rain rate and the radome diameter, see for example the Gibble's or Mei's formula in Ruze (1966). Then we determine the complex dielectric constant and loss tangent of water for the used transmission frequency and temperature. Now we calculate the multiple scattering within the film to estimate the loss and the phase shift, e.g. assuming a single homogeneous water layer. However, already Cohen and Smolski (1966) have shown that such results from theory must be verified by measurements.

The main reason for the need of verification are uncertainties about the water distribution on the radome surface, which is not uniform at all. The precipitation rate, the viscosity of the hydrometeors, the direction and the velocity of the wind, the angle of incidence, the slope of the radome surface, as well as the micro-physical properties of the surface contribute to high variability within the water cover. In the case of a hygroscopic (filming) surface: we have variable film thickness and formation of rivulets (small streams of water); while in the case of a hygrophobic (non-filming) surface: the density, the diameter, the shape and the angle of contact of the droplets, rivulets and streaks vary. After Ruze (1966), the water film flow is comparatively slow whereas the streaks may have velocities up to the terminal fall velocity of drops. Since rivulets and streaks are mainly vertically oriented, attenuation and phase shift is dependent on polarization, which reduces the accuracy of polarimetric measurements.

What can we do to reduce radome attenuation? It has been shown that a continuous film of water on the surface of the radome is much more lossy than isolated droplets (see for example Effenberger et al., 1986; Anderson, 1975; Cohen and Smolski, 1966). Thus holding the radome surface hydrophobic (non-filming) is more efficient than reducing the dry loss by means of special materials. Furthermore keeping the surface clean results in better water run-off and thus thinner water covers which again helps to reduce the losses. Anderson (1975) has found a considerable degradation of the non-wetting surface properties of the radome of
a satellite communication antenna, both by observing the water cover as well as by measuring the transmission loss. Within the first half year of the experiment one-way attenuation at a rain rate of 10mm/h increased from below 1dB up to about 7dB (transmission frequency was 20GHz). Assuming a homogeneous water layer at 5.5GHz this corresponds to below 1dB up to about 3.5dB. After that no further degradation has been observed. This suggests that already in the first few months of operation atmospheric pollution and weathering may seriously change the wetting properties of a radome.

4.4 Radome Attenuation under Natural Conditions

To estimate radome attenuation using real data we compare the echoes of precipitation systems observed during both, a period of a dry and a wet radome. Already a simple comparison of the average reflectivity will reveal the phenomenon, if significant. A quantitative estimate is achieved by carefully eliminating other influences on the radar echo during data acquisition.

To produce the complex water cover on the radome artificially is rather difficult. To wait for real precipitation at the radar site is much simpler and gives natural conditions. Then reflectivity data collected before and after it starts raining at the station can be compared.

The method requires all other factors influencing the radar echo to be separated from the influence of radome attenuation. This can be done by integrating the reflectivity measurements in a manner that the sum of the echo increases and decreases caused by other factors is small compared to the echo change caused by the water cover on the radome. Precipitation growth and decay, e.g., do not cause any problem, as long as they are slow averaged over a large precipitation area (in the following studies larger than 10'000km²). The same yields for atmospheric signal attenuation depending mainly on precipitation intensity. Often it is better to track single precipitation systems, e.g. a thunderstorm, a squall line or a front, than averaging over the whole radar image. Therefore, we define regions of interest enclosing the desired area. This avoids entering and leaving of cells to cause abrupt echo increase and decrease allowing us to compare data of the same population. The water cover on the radome may be caused by the system investigated or by a separate precipitation area.

Polar data was used for the analyses. First, because artefacts introduced by the transform from polar to cartesian are out of question. Such artefacts may be caused by anything that systematically varies with elevation angle (beam shielding, mean reflectivity, frequency of ground clutter, time, ...). Second, the slope and the exposition to the wind produce a complex dependence of the resulting attenuation from azimuth and elevation. For quantitative precipitation estimation it is interesting to know the dependence on the elevation angle.

In this study reflectivity data of elevation scans taken at angles of -0.3, 0.5, 1.5 and 2.5deg are
analysed separately. At elevations larger than 5deg, echo changes caused by the combination of a) the vertical decrease of reflectivity, b) the conical surface of the single elevation scan and c) the movement of the precipitation system relative to the radar may be of the same order of magnitude as radome attenuation.

At the radar sites Albi and Monte Lema the precipitation intensity over the radome has been retrieved from the closest reflectivity measurements. On mount La Dôle there is also a rain gauge which allows a comparison.

Reflectivity measurements within the defined regions of interest have been analysed as follows: 1) The logarithmically averaged reflectivity decrease within the precipitation area at the moment when it starts raining at the radar site reveals the attenuation in dB. 2) Linearly averaging the precipitation rate gives a quantitative measure of the resulting average underestimation (relevant for hydrologic applications). The difference between the logarithmically averaged reflectivity and the linearly averaged precipitation rate is given by the intensity probability distribution and the Z-R relationship (here $Z=316R^{1.5}$). 3) Detailed insight can be obtained by the reflectivity probability distribution (section 4.4.4). 4) A matrix has been defined for the investigation of the reflectivity pattern transformation (see section 4.4.3). 5) Finally the total area of the region of interest (roi) and the area of the precipitation system show the involved geographic extensions. The precipitation area is defined as the area within the roi with reflectivity greater than 13dBZ during a dry radome.

### 4.4.1 Investigating the Wetability of the Radome

Experiments have been made on the three-year old radome on Monte Lema in order to investigate the wetability of the Swiss radomes. The radome on Monte Lema is of the same fabric as the ones on Albi and mount La Dôle. They are operative since 1993, 1994 and 1995 respectively. The results of the experiments are the following: The first drops falling on the radome form single droplets. Soon they collapse and connect with new raindrops, forming a film of water covering the surface of the radome. In the steeper parts of the radome this process gets accelerated by gravity and to the luv of the radome by the wind. Once a film of water is formed it remains until it dries. Observing the process of drying showed that the last state of the water cover was a very thin film. A hydrophobic behaviour of the radome may be negatively affected by adhering dirt and dust.

### 4.4.2 The Cold Front on 21 May 1996

On 21 May 1996 a cold front was traversing the Jura mountains. The front was only about 60km wide and besides it the radar screen showed several separate precipitation areas. They
Figure 4.1: Cold front passing the radar station La Dôle on 21 May 1996. Elevation is -0.3deg. Logarithmically averaged reflectivity within the precipitation area and precipitation rate on mount La Dôle versus time. The gauge measured a total amount of 6.4mm, the radar 5.3mm. During precipitation at the site there is a signal decrease of at least 3.4dBZ.

were moving from west to east with a ground speed of about 40km per hour, and were thus visible from the radar for half a day (maximum range is 230km). While the front was passing the radar station on mount La Dôle the rain gauge recorded moderate rain rates, i.e. 6.4mm in 90min. Temporarily the air temperature fell below the freezing point (-0.4°C) and it is possible that a mixture of water and ice covered the radome. Figure 4.1 depicts the logarithmically averaged reflectivity within the precipitation area of 15'547km² connected to the front and the rain rate at the La Dôle site versus time. Radome attenuation is clearly revealed. Comparing data taken at 1532 and 1607 UTC leads to an estimate of at least 3.4dBZ two-way attenuation. The first non-zero signal of the rain gauge corresponds to the period between 1600 and 1610 UTC. If the decrease in reflectivity starting at 1532 UTC is explained by radome attenuation, the radome must have been wetted before. This is quite probable, since the measured relative humidity was very high. A dry, heated gauge needs 0.2mm to be wetted, at low rain rates much more. And some of the rain may have been missed because of
wind drift (average velocity around 9 ms$^{-1}$). The increase of the average reflectivity between 1707 and 1732 UTC is as abrupt as the decrease 90 min before and just at the moment when it stopped raining. This evidently substantiates the thesis of radome attenuation.

### 4.4.3 The Cold Front on 19 May 1996

Again the decrease of the logarithmically averaged reflectivity within the precipitation area has been determined. We have found signal attenuation values of 1.8 and 2.4 dBZ (see Table 4.2). For detailed insight a matrix has been calculated containing information of the transformation of the reflectivity patterns at 0647 (dry radome) and 0707 UTC (wet radome), see Table 4.1. Its element ($i$th column, $j$th row) equals to the area in km$^2$ of all pixels that belong at 0707 UTC to the reflectivity class $i$ and at 0647 UTC to the class $j$. The matrix has been calculated as follows: For each pixel the matrix element ($i,j$) was determined ($i$ is the reflectivity class at 0707 UTC and $j$ at 0647 UTC) and incremented by the area for which the pixel is representative. If the two patterns were absolutely identical the matrix would be of diagonal type. Assuming constant radome attenuation, let us say a decrease by one class, and a stationary precipitation system only the matrix elements with $j = i + 1$ would be non-zero. If the precipitation system is moving the matrix becomes smeared (the non-zero elements are no more aligned). A uniform matrix, on the other hand, tells us that the patterns are in a sense completely uncorrelated.

The matrix given in Table 4.1 has diagonal character but is asymmetrical and sort of smeared. The latter can be explained by the horizontal shift of the front of about 10 km (that is the eighth part of the west-east extension) and the movement of the cells within the front relative to each other. The asymmetry is the result of radome attenuation. Considering the regions of the same change in reflectivity, the reduction by one class (i.e. 3 dBZ) is the most frequent change. In particular within an area of 5'496 km$^2$ (sum of matrix elements in italic shape) reflectivity has diminished by 3 dB, within 3'655 km$^2$ even by 6 dB, while there has been an increase of 3 dB over an area of only 2'268 km$^2$.

The above mentioned estimate of radome attenuation of 2.4 dBZ can be obtained from the column and row sums of the matrix as follows: First estimate the precipitation area, using data before rain on the radome: $PreArea = \sum_{i=1}^{10} A_{647}(Z_i)$, where $A_{647}(Z_i)$ is the area of all pixels of the reflectivity class $Z_i$ within the ROI at 0647 UTC, class 1 is the class 13–16 dBZ. The decrease of logarithmically averaged reflectivity within the precipitation area is then $\frac{1}{PreArea} \{ \sum_{i=0}^{10} [A_{647}(Z_i) - A_{707}(Z_i)]Z_i \}$ where $Z$ is in dBZ and $Z_0$ is set to 11.5 dBZ. Note that for the events in July 1996 and later (listed in Table 4.2) both data before and after rain on the radome have been considered, which allowed to estimate the mean echo growth or decay during the measurement.
4.4.4 The Occluded Front on 14 July 1997

Figure 4.2 depicts the reflectivity probability distribution (grey scale), the logarithmically averaged reflectivity, the linearly averaged precipitation intensity, as well as the rain rate at the radar site of an occluded front on 14 July 1997. The rain at the site was of very short duration. The Log-AVE reflectivity varied very weakly, except during the rain on the radome. Similarly, the probability distribution shows a smooth overall precipitation growth and decay interrupted by an abrupt change caused by the attenuation due to local rain. The estimates of radome attenuation can be found in Table 4.2.

4.5 Discussion

Seven precipitation events have been analysed. In two of them the influence of radome attenuation was covered by other effects and could therefore not be isolated and estimated.
4.5. Discussion

Table 4.1: Reflectivity transformation matrix, cold front observed from La Dôle on 19 May 1996. Elevation is 1.5deg. The matrix element (ith column, jth row) denotes the area in km² over which reflectivity changed from class j at 0647 UTC (dry radome) to class i at 0707 UTC (wet radome). The back column and the bottom row hold the sums. See section 4.4.3.

<table>
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<th>&lt;13</th>
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<th>&lt;19</th>
<th>&lt;22</th>
<th>&lt;25</th>
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<th>&lt;31</th>
<th>&lt;34</th>
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<td>135</td>
<td>23</td>
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<td>0</td>
<td>166136</td>
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Table 4.2: Results of the analysed precipitation systems.

<table>
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<tr>
<th>Radar site</th>
<th>Date</th>
<th>Time UTC</th>
<th>roi km²</th>
<th>PreArea km²</th>
<th>R mm/h</th>
<th>△LogAVE-Z dBZ</th>
<th>△LinAVE-R dBR</th>
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<tbody>
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<td>Albis</td>
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<td>2235</td>
<td>50'504</td>
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<td>5.0</td>
<td>3.7</td>
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<td>1830</td>
<td>61'236</td>
<td>10'254</td>
<td>12</td>
<td>3.0</td>
<td>2.6</td>
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<tr>
<td>La Dôle</td>
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<td>1010</td>
<td>15'547</td>
<td>6.6</td>
<td>3.4</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Monte Lema</td>
<td>24Jul96</td>
<td>0645</td>
<td>10'394</td>
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<td>-</td>
<td>2.6</td>
<td>1.3</td>
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<td>0710</td>
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<td>-</td>
<td>1.8</td>
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<td>23'449</td>
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<td>0.8</td>
<td>0.7</td>
</tr>
</tbody>
</table>

*Time of decrease measurement. (Subtract 3 minutes for elevation scans at -0.3 and 1.5deg)
*Observed region of interest
*Precipitation area at elevation 0.5deg (except 21May96: -0.3deg) at the time of the decrease measurement. Estimated from area with reflectivity >13dBZ before and after rain on the radome (except May96: area >13dBZ before rain on the radome).
*Precipitation rate at the radar site at the time of the decrease measurement. La Dôle: gauge, Albis and Monte Lema: radar.
*Decrease of the logarithmically averaged reflectivity within the precipitation area at elevation angles of -0.3, 0.5, 1.5 and 2.5deg
*Decrease of the linearly averaged precipitation rate, at elevation 0.5deg (except 21May96: -0.3deg).
For the same reason some estimates are lacking on 24 July 1996. The results of the remaining five are summarized in Table 4.2. The archive of polar radar data is operational since summer 1996, thus the gaps in the event May 1996. On 2 July 1997 two measurements have been made, one at moderate-to-strong and the other at small rain rates. Attenuation estimates obtained by means of the proposed technique have an estimated error of the order of ±0.5dB. Largest attenuation numbers have been found on 14 July 1997: 5.4dB two-way attenuation of reflectivity in moderate rain, which led to an average underestimation of the rain rate of 3.7dB (only 43% of the rain was measured).

To compare the measurements with theory we determine the equivalent film thickness, i.e. the thickness of a homogeneous water film that introduces the same attenuation as measured. According to Joss (1996a) we obtain values between 0.03 and 0.14mm. Using Gibble’s approximative formula (see Ruze, 1966) for the given rain rates the thickness ranges from 0.05 to 0.12mm. We conclude that the measured attenuation numbers have the same order of magnitude than values expected from theory.

Obviously radome attenuation can significantly reduce the accuracy of quantitative precipitation estimates. The number of studied cases is too small to investigate the relation between radome attenuation and the rain rate or the elevation angle. Both from theoretical considerations and from the experience with real data that has been gained so far, we do not expect a simple relationship and thus conclude that correction of radome attenuation will be a difficult task.
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References


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Curriculum Vitae

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born in Trogen AR, Switzerland, on 8 December, 1971

Education

1978–1985 Primary and secondary school in Trogen AR
1986–1991 High school in Trogen AR
Matura: Typus B, Jan. 1991
Nov. 1992–Apr. 1997 Department of Geography, University of Basel
Examiner: Prof. E. Parlow, advisors: G. Galli and Dr. J. Joss
Sep. 1993–Dec. 1995 Junior Scientist at the MCR-Lab, University of Basel
Operating airborne infrared camera, advisor: Dr. C. Roggo
Jun. 1995 Landeshydrologie und -geologie, Bern
Practical training, advisor: Dr. B. Schädler
Sep.–Oct. 1995 NORUT Informasjonsteknologi, Tromsø, Norway
Lasfin — a digital image processing module for large-scale forest
inventory (written in C), advisor: Dr. A. K. Gjertsen
Define and implement algorithm for operational wind profile product
Modify decision-tree for clutter elimination
Improve real-time precipitation estimates in the Alps
Oct. 1997–Nov. 2000 Ph.D. student at the Institute for Atmospheric Science, ETH, Zürich
Examiner: Prof. A. Waldvogel, co-examiner: Dr. J. Joss
Aug.–Nov. 1999 Mesoscale Alpine Programme (MAP):
Radar scientist at the Project Operations Centre in Milano
and at Locarno-Monti (SOP)
Instrument intercomparison experiment (Pre-SOP)
May 2001– Post-doctorate at the McGill University, Montréal, advisor: Prof. I. Zawadzki
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             (with Dr. J. Joss and Prof. K. Beheng)
Mar. 2000    One-week course with meteorological excercises and problems
             University of Basel
2000        Lectures on 'Meteorological Risks in the Mountains'
             for scouts, Jugend-Sport guides, and teachers

International Conferences

Mar. 1998    COST-75 Final International Seminar, Locarno, Switzerland
             Oral presentation: Profile-correction in the Alps
Jun. 1998    MAP Meeting 1998, Chamonix Mont-Blanc, France
             Oral presentation: New Robust De-Aliasing for Vertical Wind Profile
Sep. 1998    25th International Conference on Alpine Meteorology, Torino, Italy
             Oral presentation: Variability of Alpine Precipitation
Apr. 1999    EGS General Assembly 1999, The Hague, Netherlands
             Poster: Clutter Suppression Again?
Jun. 1999    MAP Meeting 1999, Appenzell, Switzerland
             Poster: Profile-correction in the Alps
Jul. 1999    29th Conference on Radar Meteorology, Montréal, Canada
             Oral presentation: Profile-correction in the Alps
May 2000    MAP Meeting 2000, Bohinjska Bistrica, Slovenia
            Oral presentation: Spatial Continuity of Alpine Precipitation
Sep. 2000    1st European Conference on Radar Meteorology, Bologna, Italy
            Keynote: Spatial Continuity of Alpine Precipitation

Intragna, 20 November 2000

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List of Publications


Acknowledgment

It all started in November 1995 with a letter to Dr. Jürg Joss, head of MeteoSvizzera in Locarno-Monti. At that time, I was looking for a subject and an advisor of my diploma thesis. Five years later, still in Locarno-Monti, I am writing the last lines of my Ph.D. thesis, and thinking back to my first discussion with Jürg and to my first contacts with radar meteorology. Many open conversations followed in the morning-train from Intragna to Locarno or in the office. I most appreciated his thoroughness, creativity, and infinite interest in almost any subject. It was an exciting time and very instructive too. Thank you.

I would like to express my gratitude to Prof. Albert Waldvogel, the examiner of this Ph.D. thesis. Although he left the Institute for Atmospheric Science to become Vice-President of the ETH for Research and Economic Relations a few months after I started my Ph.D., he always supported my work and helped me to finish the thesis in time.

My particular gratitude goes to the colleagues in Locarno-Monti for the good working atmosphere, for fruitful discussions and support, for the Italian lessons, for UNIX support, and for giving insight into weather forecasting in the Southern Alps: Gianmario Galli, Remo Cavalli, Marco Boscacci, Greta Fornera, Evio Tognini, Thérèse Tiberi, Hanspeter Roesli, Paolo Ambrosetti, Giovanni Kappenberger, Fosco Spinedi, Guido Della Bruna, Laura Moser-Banfi, Elena Altoni, Claudia Spinola, Valerio Ortelli and Sergio Sartori. I would also like to thank Dr. Thomas Gutermann, former head of MeteoSwiss, who often encouraged and supported Ph.D. students working at the national weather service.

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