Doctoral Thesis

Microphysical properties of the melting layer

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Microphysical Properties of the Melting Layer

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DOCTOR OF NATURAL SCIENCE

presented by
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1998
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Abstract

Widespread precipitation, usually termed stratiform precipitation, accounts for the major part of the annual rainfall in the midlatitudes and for a large part of the precipitation in the Tropics. The physical relevant processes governing the formation of stratiform precipitation per se are well known, but to validate model calculations, observations in the field and/or the laboratory are necessary.

This thesis deals with in-situ measurements of hydrometeors in stratiform precipitation. Measurements were performed at different altitudes along the steep slope of a mountain. The two or three observing stations were chosen such that snowflakes above and/or within the melting layer could be sampled in conjunction with raindrops below the melting layer. The bottom station was equipped with a vertically pointing X-band Doppler radar and a disdrometer. At the top and if present at the middle station an optical instrument was placed to measure hydrometeors. At all stations meteorological instruments were present. The instrumentation used to measure snowflakes is capable to record particle number, size and shape. In addition, since the instruments had a fixed position on the ground, snowflake fall velocities could also be measured. The reflectivity profiles of the vertically pointing radar were used to determine the position of the optical instruments in relation to the melting layer.

The experiments were performed during the winter months of 94/95 - 96/97 with the idea to measure precipitation, uniform over a long time and range. The fixed setup of the instruments and the wish to measure vertical profiles of hydrometeor properties required a meteorological situation with a front passing the experimental site. A classification of the different precipitation events was performed to distinguish uniform stratiform precipitation from stratiform precipitation with embedded showers and convective precipitation. This was necessary to isolate the cases for which measurements at the different stations could be compared. Out of the large data set obtained during the several years, four cases could be selected to be included in this study. Two cases consist of a complete passage of the melting layer at a station and two cases consist of a partial passage.

Number fluxes of snowflakes and raindrops were calculated and used to tackle the question to what extent aggregation is active above and within the melting layer and whether breakup of melting snowflakes does occur within the melting layer. The results show that aggregation is active not only above the melting layer but also within. Qualitative estimation shows that
the efficiency of aggregation increases as the snowflakes enter the melting layer. This leads to very large snowflakes whereby the largest ones are found at temperatures between 1 and 2°C. This corresponds approx. to the height of the radar maximum. In the lower part of the melting layer, sizes decrease rapidly. This is a consequence of the collapsing ice frame within the melting flakes, but, as it can be shown, strong evidence is found that breakup of melting flakes contributes to the decrease of particle sizes. Furthermore, by comparing particle fluxes at the same time right above and below the melting layer it seems that one snowflake yields one raindrop, independent of what is happening within the melting layer.

Another open question concerning the melting layer are melt distances of snowflakes of different mass. Up to now, answers to this question could only be found either in the laboratory or by model calculations. No field experiments were available to validate these results. By comparing the measured fall velocity of melting snowflakes with the known fall velocity of raindrops, melt distances can be deduced for flakes with different masses. The fall distances obtained in these field experiments are in good agreement with recent model calculations and laboratory experiments.
Zusammenfassung


partiellen Durchgang.


Chapter 1

Introduction

The most beautiful feature of snowfall are the large snowflakes that can be observed from time to time. These snowflakes consist of hundreds of individually grown ice crystals, stuck together during their fall from far above. The mechanism which is responsible for the formation of snowflakes is called aggregation. Although aggregation of snowflakes has been observed for a long time, the actually underlying mechanism is extremely complex and difficult. It turns out that aggregation is the result of two individual processes. At first, the snow crystals or snowflakes have to collide and subsequently they have to stick together. The condition of collision is complex since the ice crystals and snowflakes are found to exercise a spinning, helical as well as a shaking or swinging motion. The condition of sticking, on the other hand, depends on the form of the ice crystals and the thickness of the quasi-liquid layer on ice, both varying strongly with temperature.

Large snowflakes are not only confined to occur in wintertime in the midlatitudes, but are found in all seasons and latitudes whenever stratiform clouds at subzero temperatures occur.

1.1 Stratiform precipitation

Stratiform precipitation was once generally thought to be related to fronts in midlatitude baroclinic cyclones where precipitation particles are allowed to form in a deep nimbostratus cloud layer. However, today it is known that stratiform precipitation can also occur in the Tropics (Houze, 1997), even if no baroclinic cyclones exist there. Stratiform precipitation accounts for the major portion of the annual rainfall in the midlatitudes and for a large amount of tropical rainfall.

In stratiform precipitation, raindrops have started their life as ice crystals within the upper region of the nimbostratus cloud layer as shown in Fig. 1.1. Tiny ice crystals, formed by heterogeneous nucleation on ice nuclei, grow by vapor deposition to form large crystals. The water vapor needed for their growth is supplied by small supercooled water droplets.
Introduction

Fig. 1.1: *Characteristics of stratiform precipitation (from Houze, 1981)*

Liquid water droplets can exist below 0°C without freezing and are sometimes observed at temperatures as low as -30°C. In the region of the cloud top of a nimbostratus cloud, temperatures are typically between -20°C and -10°C (Hobbs and Rangno, 1985). Here, the liquid water present is relatively low (up to 0.1 g/m³, Barthazy et al., 1997). From the Clausius-Clapeyron equation follows that the saturation vapor pressure with respect to a plane water surface is larger than that with respect to an ice surface. This is a result of the different value of latent heat of evaporation (liquid → gas) and of sublimation (solid → gas). If liquid water droplets and ice crystals coexist within a cloud, the vapor pressure will be such that supersaturation with respect to ice and undersaturation with respect to water is achieved. Under this condition, the ice crystals grow, according to the Bergeron-Findeisen process, by vapor deposition while the water droplets evaporate. To sustain the growth process of ice particles, humidity has to be constantly transported from below by updraft. The main criterion for stratiform precipitation is that this upward air motion is weak enough to allow the particles to drift downward while they grow (Houze, 1997). This limits this so-called mesoscale updraft to be of the order of a few tens of centimeter per second.

When the falling ice crystals reach the temperature regime of approx. -16°C, aggregation and riming become active. Riming occurs when ice crystals collide with supercooled water droplets which freeze upon collision on the surface of the ice crystals and increase their mass. However, riming is of minor importance when updraft velocities are small. The main growth process of the ice crystals is then aggregation which leads to growth of size but not to an increase of mass. The increase of mass of the ice crystals and snowflakes is still mainly due to vapor deposition.

Upon reaching the temperature level of approx. -6°C, aggregation is enhanced and subsequently large snowflakes are formed. The enhanced efficiency of aggregation is related to the quasi-liquid layer forming on the surface of ice (for details see Pruppacher and Klett, 1997).

When large snowflakes fall below the 0°C isotherm, they begin to melt. Since no strong updraft is present, melting takes place within a confined vertical range, forming a well defined melting layer. This behaviour contrasts precipitation accompanied by heavy convection where unmelted particles (graupel, hail) can be observed at any height and temperature.
as well as large raindrops high above the 0°C level, implying that melting does not take place at a defined height. The vertical extent of the melting layer in stratiform precipitation depends on several parameters such as the rainrate, the lapse rate or the humidity. Within the melting layer, snowflakes begin to melt and the liquid water is distributed within the crystal frame (Knight, 1979, Fujiyoshi, 1986). Aggregation is still possible within the upper part of the melting layer, whereas breakup of melting particles may occur within the whole melting layer. During melting the shape and fall behaviour of the snowflakes changes and their fall velocity increases rapidly (Mitra et al., 1990).

1.1.1 Radar observations of stratiform precipitation

The main features of stratiform precipitation, the well defined melting layer and the horizontal structures (hence the name "stratiform"), can easily be identified with radar. In a vertical cross section of stratiform precipitation, radar reflectivity increases from cloud top to the melting layer. The melting layer is seen by a radar as a region of increased reflectivity with values higher than above or below the melting layer. This region is called "bright band", referring to the bright band seen on monochromatic radar displays in earlier times when real time color coding of data was not available and radar reflectivities were displayed with different intensities. Soon after the bright band was discovered, its relation to the melting layer was recognized (see Atlas and Ulbrich, 1990 and references therein) and subsequently the relevant quantities in the formation of the bright band could be identified: important factors are the dielectric constant of the melting flake, the distribution of the melted water within the flake, the fall velocity of flakes and hence of the concentration of the flakes within space and the orientation of nonspheric snowflakes.

1.1.2 Hydrometeor size distribution

The size distribution of raindrops has been investigated by researchers with various methods for at least one century. One of the first functional relation of raindrop number concentration to raindrop diameter was presented by Marshall and Palmer (1948). The authors showed that the concentration of raindrops within the atmosphere in dependance of their diameter could be approximated by an exponential relation. This exponential relation, characterized by an intercept parameter $N_0$ and a slope parameter $\Lambda$, was found to be dependent only on the rainrate, whereby the parameter $N_0$ was found to be constant and the parameter $\Lambda$ a function of the rainrate. The concept of an exponential size distribution was expanded by Ulbrich and Atlas (1984) to the so-called gamma-distribution which includes the exponential size distribution as a special case. Furthermore, it was shown by Waldvogel (1974) that the intercept parameter $N_0$, which was proposed by Marshall and Palmer to be constant, could change considerably within short periods, not only when the rainrate changed, but also when
it remained constant.

As the exponential shape of raindrop size distributions was recognized, snowflake size distributions were investigated. They were found to obey also an exponential distribution, independent whether their real diameter was observed (Passarelli, 1978, Lo and Passarelli, 1982), or the diameter of the resulting melted drop (Gunn and Marshall, 1958).

1.1.3 Studies related to stratiform precipitation

In-situ measurements of hydrometeors in stratiform precipitation

To investigate snowfall and the melting layer, experiments were performed where snowflake size distributions above the melting layer and alternatively raindrop size distribution below the melting layer were sampled at the same time. Two main experimental setups are possible to sample snowflakes above the melting layer in conjunction with raindrops below the melting layer: Either airplanes are used or a steep mountain is needed with a top station above the melting layer. Both experimental setups were used successfully. Ohtake (1969, 1970) made extensive measurements of raindrop and snowflake size distributions at three stations along the slope of Mt. Zao in Japan. He concluded that no aggregation or breakup of hydrometeors takes place within the melting layer and that the size distribution of raindrops is dependent on that of the precursor snowflakes above the melting layer. Another set of experimental data regarding hydrometeor size distributions are reported by Yokoyama et al. (1985). The authors made along Mt. Fuji similar measurements as Ohtake. They concluded that aggregation is effective within the upper part of the melting layer and that breakup may have an effect within the lower part of the melting layer.

Aircraft measurements were done by Passarelli (1978) to obtain snow size spectra at different levels. These experimental data were used to calculate aggregation efficiencies for different type of ice crystals. The author finds that the aggregation efficiency of dendritic crystals and aggregates is greater than unity for the case study investigated. As he concludes, this may be due to wake capture. The efficiency of aggregation calculated in his paper applies for the temperature range between -12 and -15°C. No aggregation efficiencies are given for other temperature regimes, e.g. near the 0°C level. More results of aircraft data are reported by Lo and Passarelli (1982). The authors suggest that snow evolves through at least three stages characterized by deposition, aggregation and breakup. The breakup process serves to limit the number of large snow particles and interacts with aggregation to produce a limiting value of the slope of the snowsize spectrum. However, this study is also confined to heights above the 0°C level.

Aircraft measurements of the melting layer can be found in Willis and Heymsfield (1989) where experimental data are presented measured within the stratiform part of a mesoscale convective system. An isothermal layer below the 0°C level is reported within which the
major part of the mass is melted. During melting, very large aggregates are produced which can be accounted for by an increase in the terminal velocity difference between similar-sized hydrometeors, which results from differing degrees of melting.

Radar observations of the melting layer

Although in-situ measurements are important, considerable insight into the microphysical properties of the melting layer are possible by radar observations only. Yokoyama et al. (1984) present observations of the melting layer with two radars at X- and C-band. Aggregation of hydrometeors within the melting layer is detectable, especially in the case of stationary precipitation. In the case of precipitation with cellular structures less evidence for aggregation is found. Furthermore, evaporation within the melting layer can be observed when the atmosphere is dry around the 0°C level.

An UHF windprofiler with Doppler capability is used by Drummond et al. (1996) to investigate the melting layer. The authors suggest that aggregation is occurring much of the time in the melting layer but that breakup effects become dominant in heavy precipitation. Longterm studies with the same windprofiler and a vertically pointing X-band radar are reported by Fabry and Zawadzki (1995). The classical explanation of the bright band is found to account for less than half of the observed reflectivity enhancement. The authors suppose that the difference could be explained by effects associated with the shape and density of melting snowflakes and, to a smaller extent, by precipitation growth in the melting layer and aggregation in the early stages of the melting followed by breakup in the final stage.

Theoretical modeling of stratiform precipitation

Stratiform precipitation and especially the microphysical processes within the melting layer can be investigated with theoretical models. Since stratiform precipitation shows not much horizontal variability, one-dimensional models can be used. Klaassen (1988) presents such a model that calculates the evolution of the particle size distribution which is used in turn to calculate radar reflectivities. Model outputs are validated with vertical profiles of radar reflectivity and Doppler velocity. The authors state that the reflectivity calculated by the model is very sensitive to the density of the melting particles and to the dielectric properties, i.e. to the amount of liquid water within the melting snowflakes. The influence of aggregation is reported to be less important.

Another interesting quantity within the melting layer is the melt distance of snowflakes. Matsuo and Sasyo (1981) calculate melt distances for snowflakes of different size and density, melting in air with different humidity. They show that at $f = 80\%$ it takes the largest snowflakes (melted diameter 5 mm) approx. 700 m to melt completely. This is reported to correspond to the observed maximum width of the radar bright band.
At last, theoretical model results can also be compared to laboratory experiments. At first thought, it seems to be impossible to reproduce such a complex process as precipitation under laboratory conditions. But certain aspects of the microphysics of precipitation can be studied very well in the laboratory, e.g. in a wind tunnel. Such wind tunnel experiments with melting snowflakes are presented by Mitra et al. (1990). The authors recorded by cinematography the variation of the fall mode of snowflakes, their fall velocity and the percentage of ice melted as a function of the travelled distance. The laboratory results were used to validate a theoretical heat transfer model. Model and observations yielded reportedly melt distances for snowflakes with melted diameters of 2.7 mm of 350 to 600 m.

1.2 State of the art

The melting layer of stratiform precipitation is a wide field of research with a large amount of work done by several generations of researchers. A short glimpse at this work was given in the previous section. The following list summarizes some of the general findings:

- Aggregation of hydrometeors occurs not only above the melting layer but also within the melting layer.
- Whether or not breakup of hydrometeors occurs within the melting layer is controversial.
- To explain the increased radar reflectivity at the level of the melting layer, the shape and density of snowflakes seems to play a crucial role.
- Melting snowflakes exercise unpredictable motions during their fall.
- The distribution of melt water within a melting snowflake is fairly well known.
- Although the actual melting process of snowflakes is complex, melt distances can be calculated rather reliably.

1.3 Open questions

A study of present literature shows that the main problem in investigating the melting layer is not the lack of knowledge of the physical relevant processes taking place within the melting layer, but the lack of detailed in-situ measurements of the relevant parameters which enter the theoretical models. This involves a detailed in-situ study of aggregation as well as breakup—if latter occurs at all. It furthermore involves in-situ measurements of melt distances, the range of fall velocities of the snowflakes, the frequency of collisions and the shapes and orientations of the snowflakes.
Problems concerning in-situ measurements of hydrometeors

In-situ measurements of hydrometeors in precipitation and especially within the melting layer have intrinsic problems that are not easy to overcome. First of all, if a vertical profile of the microphysical properties through the melting layer is desired, either the observer or the melting layer has to move. The melting layer descends or ascends under certain meteorological conditions and this can be used to measure a profile through the melting layer from a fixed, ground-based station on a mountain. But since precise weather forecasts are difficult and high mountains are not always within reach, ground-based measurements are not popular. On the other hand, airborne missions have the great advantage of being free in following the height of the melting layer. However, they have the disadvantage that no continuous vertical profile can be flown. At best, continuous data may be obtained along loops flown with a diameter of several kilometers, or discreet measurements can be made along the same sector of each loop at different heights. But still, the fall velocity of hydrometeors cannot be measured with airborne instruments.

The instrumentation needed may also impose some problems. To investigate snowflakes, optical instruments are required. Such instruments are widely used, but they are designed for airborne use only. Since only ground-based instruments are capable of measuring the fall velocity of ice crystals and snowflakes, instruments have to be developed for the use at ground station.

1.4 The motivation of this thesis

The presently open questions concerning the microphysical processes within the melting layer and the problems concerning field experiments were summarized in the previous section. Of course, not all questions may be answered in this thesis. However, a few important questions will be tackled. Central to the present study was a new instrument, the optical spectrometer. With this instrument, continuous ground-based measurements of snowflake sizes and velocities could be made. Furthermore, the unique advantage of having a variety of mountains with steep slopes which could easily be reached from the institute suggested to carry out in-situ ground-based measurements of stratiform precipitation in the tradition of the Japanese researchers, combined with radar measurements.

The thesis is divided into six chapters:

Chapter 2 gives an overview of the instrumentation used in the field experiments. The major part of this chapter is devoted to the description of the optical spectrometer, developed at the ETH.

In Chapter 3 a case study is presented based on assumptions concerning the microphysics within the melting layer.
Chapter 4 introduces some more experimental data, obtained in a measuring campaign running for five month in the winter of 96/97.

The assumptions stated in Chapter 3 are tested in Chapter 5 with additional cases. All cases together are then looked at under some guiding principles.

Chapter 6 presents the conclusions and Chapter 7 some concrete plans for further field experiments as well as suggestions for future research.
Chapter 2

Instrumentation

This thesis is based upon data obtained in several field experiments. In this chapter the most important instrumentation used for the experiments is introduced. These are two radars, two optical particle measuring instruments and a disdrometer. The properties of the radars are shortly summarized without giving a detailed overview on the principles of radar measurements. However, additional information can be found in radar books, e.g. in the witty book of Rinehart (1991) or in the book of Atlas (1990). One of the particle measuring instruments was developed at the ETH and is therefore discussed very detailed. The other was used in cooperation with the Joanneum Research Institute in Graz, Austria and its principle is shortly summarized. At last, a few words are added about the disdrometer.

2.1 Radars

Two radars have been used, one mobile and vertically pointing X-band Doppler radar and one scanning C-band Doppler radar with fixed position.

2.1.1 The mobile X-band Doppler radar

The vertically pointing Doppler radar (Mosimann et al., 1993) is mounted on a van (see Fig. 2.1). It has a wavelength of 3.2 cm and a pulse-repetition frequency of 4000 Hz. The peak power is 90 kW and the pulse length is 0.2 μs which corresponds to a distance of 30 m, but vertical resolution is set to 50 m. Beam angle is 2.4°. The range of Doppler velocities (Nyquist-interval) is $|v| = 32$ m/s and the resolution was initially 128 pulses per range gate, i.e. $Δv = 0.5$ m/s. Velocity resolution was increased in the summer of 1996 to 512 pulses per range gate, i.e. $Δv = 0.125$ m/s, in order to have better measurements of fall velocities of snow. Time resolution is variable, usually 30 s was used. The raw data are processed and yield a full Doppler spectrum for each range gate out of which several moments can
be calculated such as radar reflectivity $Z$, Doppler velocity $v_m$ or the width of the Doppler spectrum $\sigma_v$.

Reflectivity and Doppler velocity data of the X-band radar are usually depicted in a height-time-diagram (HTI) where the horizontal axis represents the time and the vertical axis the height. The vertical profiles are printed subsequently next to the other and thus yield a two-dimensional picture.

### 2.1.2 The C-band Doppler radar

The C-band radar (Li et al., 1995) is mounted on the roof of the Institute of Atmospheric Science, ETH, in Zürich. It is used to survey precipitation by recording PPIs and RHIs. A PPI (Plan-position indicator) is a display on which radar reflectivities are shown in plan position with range and azimuth angle displayed in polar coordinates, forming a map-like picture. An RHI (Range-height indicator) is a display with the height as the vertical axis and range as the horizontal axis.

Properties of both, the X-band and the C-band radars are summarized in Tab. 2.1.
### Table 2.1: Properties of the C- and X-band radars used for these case studies.

<table>
<thead>
<tr>
<th></th>
<th>X-band (mobile)</th>
<th>C-band</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wavelength (cm)</td>
<td>3.20</td>
<td>5.33</td>
</tr>
<tr>
<td>Peak transmitted power (kW)</td>
<td>90</td>
<td>250</td>
</tr>
<tr>
<td>Beam width circular (deg)</td>
<td>2.4</td>
<td>1.6</td>
</tr>
<tr>
<td>PRF (kHz)</td>
<td>4.0</td>
<td>0.25–1.25</td>
</tr>
<tr>
<td>Pulse length (m)</td>
<td>30</td>
<td>75–450</td>
</tr>
<tr>
<td>Antenna rotation (deg/sec)</td>
<td>–</td>
<td>18</td>
</tr>
<tr>
<td>Polarization</td>
<td>linear</td>
<td>linear horizontal</td>
</tr>
<tr>
<td>Operation mode</td>
<td>vertically looking</td>
<td>PPI, RHI sector scan</td>
</tr>
<tr>
<td>Transmitter</td>
<td>Magnetron type</td>
<td>Magnetron type</td>
</tr>
<tr>
<td>Receiver:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Reflectivity channel</td>
<td>logarithmic</td>
<td>logarithmic</td>
</tr>
<tr>
<td>Doppler channel</td>
<td>Hard limiting</td>
<td>linear with STALO</td>
</tr>
<tr>
<td>Doppler filters</td>
<td>–</td>
<td>and COHO</td>
</tr>
<tr>
<td></td>
<td>14 infinite impulse response</td>
<td>with varying stopband and passband</td>
</tr>
</tbody>
</table>

### 2.2 The optical spectrometer

The optical spectrometer (Fig. 2.2) was developed and built at the Institute of Atmospheric Science at the ETH in Zürich. A light source (an incandescent lamp) illuminates a one-
12 Instrumentation

dimensional horizontal array of photoelements, oriented normal to the light beam. The photoelements are read with a frequency of 7.3 kHz. The array consists of 256 elements and yields a horizontal resolution of 0.146 mm. Therefore, minimum particle size is 0.146 mm. The maximum particle size is 33.9 mm which is the mean diameter of the largest size class. Each particle falling through the measuring plane casts a shadow on the array. The consecutive data of this array yield a pseudo-two-dimensional image of the hydrometeor. This image consists of a number of rows depending on the fall velocity and size of the particle and the read-out frequency. Thus, the vertical resolution of the image depends on the fall velocity of the particle. The horizontal resolution of the optical spectrometer is not good enough to identify different crystal types. However, if single crystals are large and conditions are ideal, they can be identified like the 3 dendrites depicted in Fig. 2.3.

![Fig. 2.3: Optical spectrometer images of three perfect dendrites.](image)

The maximum measuring plane is a rectangle of the size 37.23 × 100 mm², whereby the beamwidth of the light source is 37.23 mm. Particles falling through the edges of the measuring plane are rejected if the following condition is fulfilled: The number of rows of the image touching one edge is larger than the total number of rows divided by two. This condition leads to a measuring plane the size of which depends on the size of the measured particles. Tab. 2.2 shows that the larger the mean diameter of a size class is, the smaller is the active measuring area. The image and the exact recording time of each particle are stored and processed later. The evaluation software determines the diameter of a particle as the length of the row with the largest number of covered elements. Particle images are sized into 25 size classes.

The sample volume of the optical spectrometer depends on the particle fall velocity and the particle size. The measuring area of size class 1 of the optical spectrometer is 3716 mm² compared to 5030 mm² of a Joss-Waldvogel disdrometer. The measuring area of the optical spectrometer decreases by only 6% for raindrops up to 4 mm which might be a reasonable upper limit for stratiform precipitation. Therefore, if both instruments measure stratiform rain, the sample volume of the optical spectrometer is by about 30% smaller than the sample volume of the disdrometer. If snow is measured with the optical spectrometer, the sample volume decreases further according to the lower fall velocity and larger sizes of snow particles. For a 10 mm snowflake with a fall velocity of approx. 2 m/s and the melted diameter of 2.7
Table 2.2: Mean diameter, class width and measuring area of the 25 size classes of the optical spectrometer.

<table>
<thead>
<tr>
<th>Class</th>
<th>Diameter (mm)</th>
<th>Width (mm)</th>
<th>Area (mm²)</th>
<th>Class</th>
<th>Diameter (mm)</th>
<th>Width (mm)</th>
<th>Area (mm²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.146</td>
<td>0.146</td>
<td>3716</td>
<td>14</td>
<td>3.942</td>
<td>0.730</td>
<td>3515</td>
</tr>
<tr>
<td>2</td>
<td>0.292</td>
<td>0.146</td>
<td>3709</td>
<td>15</td>
<td>4.745</td>
<td>0.876</td>
<td>3469</td>
</tr>
<tr>
<td>3</td>
<td>0.438</td>
<td>0.146</td>
<td>3701</td>
<td>16</td>
<td>5.767</td>
<td>1.168</td>
<td>3409</td>
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<td>4</td>
<td>0.584</td>
<td>0.146</td>
<td>3694</td>
<td>17</td>
<td>7.008</td>
<td>1.314</td>
<td>3334</td>
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<td>5</td>
<td>0.730</td>
<td>0.146</td>
<td>3686</td>
<td>18</td>
<td>8.468</td>
<td>1.606</td>
<td>3243</td>
</tr>
<tr>
<td>6</td>
<td>0.876</td>
<td>0.146</td>
<td>3679</td>
<td>19</td>
<td>10.293</td>
<td>2.044</td>
<td>3123</td>
</tr>
<tr>
<td>7</td>
<td>1.022</td>
<td>0.146</td>
<td>3672</td>
<td>20</td>
<td>12.556</td>
<td>2.482</td>
<td>2967</td>
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<tr>
<td>8</td>
<td>1.241</td>
<td>0.292</td>
<td>3660</td>
<td>21</td>
<td>15.330</td>
<td>3.066</td>
<td>2763</td>
</tr>
<tr>
<td>9</td>
<td>1.533</td>
<td>0.292</td>
<td>3645</td>
<td>22</td>
<td>18.688</td>
<td>3.650</td>
<td>2498</td>
</tr>
<tr>
<td>10</td>
<td>1.825</td>
<td>0.292</td>
<td>3630</td>
<td>23</td>
<td>22.776</td>
<td>4.526</td>
<td>2149</td>
</tr>
<tr>
<td>11</td>
<td>2.190</td>
<td>0.438</td>
<td>3611</td>
<td>24</td>
<td>27.813</td>
<td>5.548</td>
<td>1680</td>
</tr>
<tr>
<td>12</td>
<td>2.701</td>
<td>0.584</td>
<td>3583</td>
<td>25</td>
<td>33.945</td>
<td>6.716</td>
<td>1049</td>
</tr>
<tr>
<td>13</td>
<td>3.285</td>
<td>0.584</td>
<td>3551</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

mm the sample volume of the optical spectrometer would be by about 80% smaller than for the resulting raindrop measured with the disdrometer.

2.2.1 Determination of hydrometeor fall velocity with the optical spectrometer

To determine the fall velocity of a hydrometeor with the optical spectrometer, an assumption of the axial ratio (vertical to horizontal extent) of the particle has to be made. The simplest way is to assume this ratio to be one. If rain is measured this assumption is only valid for small drops. Large drops have axial ratios smaller than one. Since the axial ratio of raindrops depend on their size, a correction factor can easily be introduced (Beard and Chuang, 1987). However, this correction is not used, and therefore, the fall velocities for large raindrops are overestimated.

If snow is measured, an axial ratio of one might also be wrong. In order to have an idea of the axial ratio of snowflakes, a photo camera was mounted on the optical spectrometer as shown in Fig. 2.4. Pictures of freely falling snowflakes were taken about 1 cm below the measuring plane. The camera was looking in the same direction as the optics of the spectrometer, and was triggered by the optical spectrometer: if a particle larger than a certain threshold passed
the spectrometer, a picture was taken. The threshold was needed in order to control the number of pictures shot. In heavy snowfall the threshold was set to approx. 7 mm, in weak snowfall to approx. 3 mm. Fig. 2.5 gives an impression of how snowflakes look like when photographed or when recorded with the optical spectrometer.

![Snowflakes](image)

**Fig. 2.4:** Side view of the photo camera and the flash mounted on the optical spectrometer. The watch was needed to identify the optical spectrometer image of every snowflake photographed.

**Fig. 2.5:** Photograph and optical spectrometer image of three snowflakes. Note that the lowest flake consists of dendrites. Several dendrites are visible in the upper left part of the flake in the photograph. One dendrite is even visible in the optical spectrometer image. The snowflakes have sizes between 1.2 and 2.8 cm.
2.2. The optical spectrometer

Table 2.3: Properties of the four events.

<table>
<thead>
<tr>
<th></th>
<th>23 January</th>
<th>8 February</th>
<th>13 February</th>
<th>14 February</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total # of particles</td>
<td>115</td>
<td>29</td>
<td>31</td>
<td>114</td>
</tr>
<tr>
<td>Average axial ratio</td>
<td>0.74</td>
<td>0.92</td>
<td>0.94</td>
<td>1.00</td>
</tr>
<tr>
<td>Temperature</td>
<td>0°</td>
<td>-2° to -3°</td>
<td>~ 0°</td>
<td>0° to 1°</td>
</tr>
<tr>
<td>Main crystal type</td>
<td>Dendrites</td>
<td>Platelike</td>
<td>Dendrites</td>
<td>Dendrites</td>
</tr>
<tr>
<td>Wind</td>
<td>no</td>
<td>no</td>
<td>weak</td>
<td>yes</td>
</tr>
<tr>
<td>Rimming</td>
<td>lightly</td>
<td>lightly</td>
<td>lightly</td>
<td>lightly</td>
</tr>
</tbody>
</table>

A few hundred photographs from four different snowfall events were evaluated by projecting the pictures of the snowflakes onto a blackboard and measuring their horizontal and vertical extension. Ratios of height to width were calculated for each flake. The size of the snowflakes could not be determined right from the photographs since absolute dimensions were not known. Thus, the size class to which the snowflakes belong had to be extracted from the data of the optical spectrometer. Axial ratios were plotted for all snowflakes as a function of the size class but no general value or dependence could be found. The differences between the cases were too large to obtain an overall valid value. This large scatter is most probably due to the different conditions of the four cases: temperature, humidity, wind, crystal type, riming and melted fraction are believed to be important parameters. Tab. 2.3 gives an overview of the meteorological conditions and predominant crystal types of each of the four events. After the first attempt to evaluate all cases together, the cases were evaluated

![Graph](image-url)
separately (Fig. 2.6). Axial ratios still scatter considerably which is visualized by the error bars representing the standard deviations of the mean axial ratio calculated for each size class. No connection to any of the parameters listed in Tab. 2.3 seems to be obvious.

The results of the evaluation of the pictures allow a rough estimation: An axial ratio of one for snowflakes generally seems to be wrong, whereas cases may occur where it would be correct. Axial ratios smaller than 0.5 seem to be very rare. An axial ratio between 0.7 and 0.9 might represent the majority of cases quite well. Therefore, the fall velocity of the larger snowflakes (> 5 mm) is overestimated by the evaluation software of the optical spectrometer roughly by 20%. No prediction can be made about the axial ratio of the particles smaller than 5 mm. This uncertainty of the calculated fall velocity has to be considered when quantities derived from fall velocity such as particle size distribution are used.

A second possibility to estimate the error of the fall velocity introduced by using an axial ratio of one is discussed in Sect. 2.3.1.

2.2.2 Wind effects

Size and fall velocity measurements of hydrometeors with the optical spectrometer can be affected by wind in many ways. Wind-effects on measurements are certainly dependent on the direction the wind is blowing from relative to the optical spectrometer. First, the assumption is made that the horizontal component of the wind is perpendicular to the light beam. Such wind blows unobstructed through the instrument. Second, the 3-dimensional wind is separated into a vertical and a horizontal component and their influence on the optical spectrometer is investigated separately. Fig. 2.7 depicts the different influence of these wind-components on the image of a hydrometeor.

No wind: The image of a spherical hydrometeor, not disturbed by wind, is usually not a circle. The horizontal extension of the image (i.e. maximum number of covered elements, h in Fig. 2.7) is equal to the real horizontal extension of the hydrometeor. But the vertical extension of the image (v in Fig. 2.7) is dependent on the particle size, the fall velocity and the read-out frequency of the optical spectrometer. The image will be a circle when the fall velocity of the particle is approx. 1 m/s, provided that pixels are drawn as quadratic boxes. Larger fall velocities will yield an oblate, smaller fall velocities a prolate image.

Vertical wind: Vertical wind accelerates or slows down hydrometeors and fall velocities will be over- or underestimated. Only the vertical extension v of the image and hence the derived fall velocity is affected but not the horizontal extension h which determines the size of the particle (Fig. 2.7 second row left).
2.2. The optical spectrometer

Fig. 2.7: Top left: spherical hydrometeor. Top right: optical spectrometer image of the hydrometeor shown left. The vertical extension of the image depends on the vertical extension of the particle, its fall velocity and the read-out frequency of the optical spectrometer. 2nd row left: image of the hydrometeor accelerated by downward winds. 2nd row right: image of the hydrometeor affected only by horizontal winds. Bottom left: oblate particles are tilted when horizontal wind components are present. 3rd row right: the image of a tilted oblate hydrometeor affected by horizontal wind. Bottom right: the image of a tilted oblate hydrometeor affected by horizontal and vertical wind. All images: h and v give the maximum number of covered elements in horizontal and vertical direction (shaded dark grey).
**Horizontal wind**: Horizontal wind does not affect fall velocities and the determination of particle size. The image recorded with the optical spectrometer is distorted, but the height of the image and the row with the most covered elements have still the same size as without wind (Fig. 2.7 second row right).

However, if hydrometeors are oblate, they will be tilted by horizontal winds. This is one more source of errors measuring hydrometeors with the optical spectrometer. In Fig. 2.7 (third row right) the image of a tilted oblate hydrometeor is shown that is affected by horizontal wind. The size (=horizontal extension) of the hydrometeor will be underestimated and the fall velocity will be overestimated because the assumption of axial ratio equal to one is not valid any more. If vertical wind adds to the horizontal wind (Fig. 2.7 bottom right), the fall velocity will be even more overestimated because the vertical wind accelerates the particle.

Wind effects may influence the measurements with the optical spectrometer to a considerable extent, under certain conditions beyond toleration. The optical spectrometer is not suitable to measure hydrometeors, especially snowflakes, in stormy conditions. But weak winds may not alter results too much. Since the optical spectrometer was placed on a meadow (as shown in Fig. 2.2) and the light beam was less than 30 cm above ground, vertical winds are supposed not to play an important role. This is even more true since measurements were only made in cases with stratiform precipitation, i.e. no stormy conditions. Therefore, it is unnecessary to try to estimate the vertical wind that adds to the fall velocity. Horizontal winds are still possible. As it was shown, horizontal winds tilt oblate particles and thus influence the measurements. Horizontal winds may blow easily with velocities of 1 to 2 m/s 30 cm above the ground. The fall velocities of snowflakes are 1 to 2 m/s, thus the fall trajectory of snowflakes is tilted by 45° to the vertical and oblate snowflakes may be tilted by 45°. The error resulting from this tilt depends on the axial ratio of the hydrometeor which, as shown previously, can be as low as 0.5. The error in the determination of the width of a particle with the extreme axial ratio of 0.5 from a tilt by 45° is about 20%. Visual observations confirm that particle fall trajectories near the ground do not exceed a tilt of 45° to the vertical in stratiform precipitation, hence particles are not tilted more than 45°.

All considerations above are only valid if the horizontal wind component is perpendicular to the light beam. If this is not the case, horizontal wind effects may be less pronounced but new errors may be introduced, e.g. by partially shielding of the measuring plane by the housing of the instrument.

The conclusion of this subsection is that the most reliable measurements are made when no winds are present. Weak winds still can be tolerated, especially if blowing perpendicular to the light beam. Errors will then be most significant for oblate particles.
2.3 The 2D-Video-Distrometer

The 2D-Video-Distrometer is a precipitation gauge, working on the basis of video cameras. It has been developed by Joanneum Research (Graz, Austria), in cooperation with ESA/ESTEC (European Space Agency/European Space and Technology center) (Schönhuber et al., 1994, 1995). In the winter season of 96/97 a field experiment was performed at Mt. Rigi by the Institute of Atmospheric Sciences of ETH together with the Institutes of Applied Systems Technology (IAS) of Joanneum Research. The 2D-Video-Distrometer was installed on Mt. Rigi as a loan from Joanneum Research.

Fig. 2.8 gives a principle drawing of the system components of the 2D-Video-Distrometer. The instrument consists of three main parts, the sensor unit, the outdoor electronics unit and the indoor user terminal. Fig. 2.9 is a schematic drawing illustrating the operating principle of the sensor unit: The trapezoidal grey boxes are the illumination devices which in principle are extremely large-diameter optical condensors. Two line scan cameras, vertically offset from each other by approx. 6 mm, are directed towards the opening of these illumination devices. The optical system is designed in such a way that (seen through the camera lens) the slit of the illumination device appears as a relatively evenly illuminated background of extreme brightness. To the cameras any particle falling through the beam of light will appear as a dark silhouette against this bright background. In reality each of the two optical paths contains two mirrors that are used to "bend" the beam. These are omitted here to simplify the drawing. Table 2.4 gives a short summary of some of the properties of the
2D Video-Distrometer.

Table 2.4: Some properties of the 2D-Video-Distrometer (after Schönhuber et al., 1994).

<table>
<thead>
<tr>
<th>Property</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horizontal resolution</td>
<td>better than 0.22 mm</td>
</tr>
<tr>
<td>Vertical resolution</td>
<td>better than 0.3 mm</td>
</tr>
<tr>
<td>(vert. velocity &lt; 10 m/s)</td>
<td></td>
</tr>
<tr>
<td>Vertical velocity accuracy</td>
<td>better than 5%</td>
</tr>
<tr>
<td>Sampling area</td>
<td>approx. 100 x 100 mm²</td>
</tr>
<tr>
<td>Rain rate accuracy</td>
<td>better than 10%</td>
</tr>
</tbody>
</table>

The 2D-Video-Distrometer was initially designed for all kinds of hydrometeors but the evaluation software is suited best for rain and hail. Hydrometeor images are recorded with both cameras and have to be matched later. Only if the image of a hydrometeor from camera A is matched to an image from camera B the fall velocity of a particle can be derived. Image-matching is quite easy if raindrops are measured for raindrops have well defined shapes and behaviours (e.g. fall velocity-dimension relation) that allow setting up strict criteria to match the images. Images of snowflakes are much more difficult to match since snowflakes may have all kind of shapes when viewed from two directions perpendicular to each other. Furthermore, the range of fall velocities of snowflakes is much more confined than the one of raindrops and no strict relation exists between their size and fall velocity.

These difficulties make it impossible to evaluate snowflake data properly up to now. The
larger flakes (> 5 mm) can be matched reliably but not the smaller ones. Therefore, 2D-
Video-Distrometer data have been evaluated separately for the two cameras like the data of
the optical spectrometer (see Sect. 2.2). Thus, fall velocities derived from the data of the
2D-Video-Distrometer suffer from the same limitations as discussed in Sect. 2.2.1.

2.3.1 Estimation of the error of the fall velocity introduced by the
assumption of an axial ratio of one

Fig. 2.10: The fall velocity \( v \) of a snowflake is calculated assuming an axial ratio of one,
\( v_{\text{real}} \) is calculated after the images of camera A and B have been matched (Hanesch, in
preparation). \( r \) is the correlation coefficient of the linear regression.

As it was discussed in Sect. 2.2.1, snowflakes do not have axial ratios of one as a rule. Thus,
fall velocities derived with the assumption of axial ratios equal to one may be erroneous.
The 2D-Video-Distrometer offers the opportunity to measure exact fall velocities, provided
that the images of camera A and B can be matched. Although this matching does not yet
work for all sizes, snowflakes larger than 5 mm can be evaluated. For these snowflakes, the
fall velocity can be calculated in two ways. First, the fall velocity \( v \) can be deduced for one
camera with an axial ratio of one. Second, the exact fall velocity \( v_{\text{real}} \) can be deduced when
images of camera A and B are matched. Fig. 2.10 shows the difference of these two fall
velocities in dependence of the maximum width of the snowflake. The data in this figure
stem from a period of 15 minutes with stratiform precipitation on 14 December 1996. It
can be seen that fall velocities \( v \) calculated with the image of just one camera generally
overestimate the real fall velocities \( v_{\text{real}} \). The larger the snowflakes are the larger becomes
this overestimation. The assumption of an axial ratio of one is obviously not valid for large
snowflakes and would have to be corrected to a value below one. A linear fit was performed with the data in Fig. 2.10 and a significant correlation was found. The results show that the fall velocity of a 10 mm snowflake is overestimated by 0.4 m/s and of a 15 mm snowflake by 0.7 m/s for this particular case. No correction can be given for snowflakes smaller than 5 mm or larger than 15 mm.
2.4 The disdrometer

This instrument, developed by Joss and Waldvogel (1967), transforms the vertical momentum of a raindrop falling onto a measuring area to an electrical signal. With the dimension-fall velocity relation given by Atlas et al. (1973), the weight and hence the size of the sampled raindrop can be deduced. Minimum size of recorded raindrops is 0.36 mm, maximum size is 5.33 mm. Large raindrops cause oscillations of the system that lead to a dead time during which small raindrops are not sampled. This depletion of the smaller drops due to the dead time of the disdrometer was calculated and then corrected (Sauvageot and Lacaux, 1995) for all data recorded with the disdrometer.

Usually, the sampling area of the disdrometer is assumed to be constant (50.3 cm²) but in truth this is not the case. Raindrops falling on the edge of the disdrometer are not sampled correctly. A decreasing size of the sampling area would have to be considered for increasing drop sizes. The correct sampling area for a large drop with a diameter of 5 mm is reduced to 44.2 cm² which corresponds to a reduction of the original sampling area by 12%.

Since an unambiguous dimension-fall velocity relationship as well as a well defined density is needed to evaluate disdrometer data, the instrument is only suited to measure raindrops. No snowflakes or ice crystals can be measured with this instrument and as soon as the melting layer descends to the ground, results become erroneous.
Chapter 3

Case study

On 17/18 February 1995, an unusual precipitation event took place in the region of Zürich. The most striking feature of this event was the uniformity of the precipitation over a wide range (> 30 km) and a long time (> 6 h). Precipitation was triggered by a very weak cold front. No strong winds were observed nor a rapid change of any meteorological quantity. All changes occurred slowly and gradually. Although there was some variation of the rainfall intensity, this was not caused by advection of precipitation patches. Due to these conditions, the melting layer was descending slowly and steadily during the 5.5 hours of the experiment by 400 m.

Microphysical measurements were performed along the steep slope of Mt. Üetli with a vertically pointing X-band Doppler radar and a disdrometer at the base of the mountain and an optical spectrometer on top of it. At the beginning of the measurements, the melting layer was above the optical spectrometer, at the end it was located between the optical spectrometer and the X-band radar/disdrometer. The optical spectrometer at the top was first measuring rain, then melting snow and at last dry snow. The disdrometer at the bottom was measuring rain all the time.

This precipitation event will be discussed in detail in this chapter.

3.1 Experimental setup

Figs. 3.1 and 3.2 show an overview of the experimental setup. The optical spectrometer was placed on top of the mountain (top station, Mt. Üetli, 870 m ASL). A vertically pointing X-band Doppler radar (properties see Sect. 2.1.1) and the disdrometer were located 2.4 km northeast and 450 m below the optical spectrometer (bottom station, Saalsporthalle, 420 m ASL). A second Doppler radar (C-band) was located 5.7 km north (Hönggerberg, see Sect. 2.1.2) of the bottom station. It was operated in RHI and PPI modes to survey the evolution and the homogeneity of the precipitation. In addition to these two radars, the
Fig. 3.1: A cross-section of the instrumental setup at Mt. Uetli is shown. The position of the melting layer in the drawing corresponds to its position at the end of the experiment.

composit Radar data (C-band) of the Swiss Meteorological Institute (SMI) are used. These data cover an area of $500 \times 400$ km$^2$ and give a rough overview of the precipitation over Switzerland. Furthermore, the bottom station was equipped with meteorological instruments measuring temperature, humidity, pressure and wind. At the top station Formvar replicas were taken during the last few hours of the experiment after the method described by Schaefer (1956).
3.1. Experimental setup

Fig. 3.2: Map of the experimental site in Zürich. The optical spectrometer is placed on top of Mt. Üetli, the X-band radar and the disdrometer at the Saalsporthalle and the C-band radar on the building of the Atmospheric Science at ETH Hönggerberg. Reproduced with the permission of the Bundesamt für Landestopographie, 28 April 1998.
3.2 Synoptic situation

Fig. 3.4 shows the SMI surface map for 00 UTC of 17 and 18 February 1995. A weak cold front slowly passed Spain and France on 17 February and behind it cold and wet marine air was directed from northwest towards the Alps on 18 February. The weather condition in the northwest of the Alps was dominated by a flat distribution of air pressure which resulted in low windspeeds. Blocked by the Alps, the airflow produced continuous precipitation for about 12 hours. During the whole precipitation event the temperature decreased slowly and steadily due to inflow of cold air and due to airmass cooling by melting ice. Within 12 hours the temperature changed in Zürich from 8°C to -1°C. Fig. 3.3 depicts time-series of some meteorological parameters measured during the experiment at the bottom station.

Fig. 3.3: Temperature, humidity and pressure as measured during the 5.5 hours of the experiment. The gap visible in the data sets result from a short circuit at the bottom station.
Fig. 3.4: SMI surface maps of 17 and 18 February at 01:00 local time.
3.3 A classification scheme

The precipitation of the case study of 17/18 February 1995 was widespread and extremely uniform. PPIs and RHIs were recorded with the C-band radar at Hönggerberg every 10 minutes to monitor the evolution of precipitation. Fig. 3.5 shows two sets of PPIs and RHIs, where the PPIs were recorded with an elevation angle of 1.5° and the RHIs with an azimuth angle of 171°. One set was taken around 21:40 at the maximum of precipitation (4 mm/h), the other taken around 00:10 when precipitation was weak (1 mm/h). Although intensity of precipitation does vary, the main features of the precipitation pattern do not.

The circular structure of high reflectivity (more than 40 dBZ) that can be seen in the PPI at 21:40 is the bright band. The rotating radar beam transmitted with the elevation angle of 1.5° intersects the melting layer at a certain distance from the radar. The lower the melting layer is, the smaller becomes the diameter of the bright band in the PPI. Almost no bright band is visible in the PPI taken at 00:10 since the melting layer was very close to the ground.

The four pictures in Fig. 3.5 are representative for the whole case study. To prove the validity of this statement, a classification scheme was developed.

3.3.1 Criteria to be met

The question here is not to find a criterion to discriminate between stratiform and convective precipitation but to find, on the basis of radar observations, a criterion to distinguish uniform precipitation (in space and time) from other stratiform precipitation types.

When observing stratiform precipitation with radar, the most prominent feature is the bright band produced by microphysical processes occurring within the well defined melting layer. Convective, and especially deep convective precipitation lacks a well defined melting layer and hence no bright band is detectable with radar.

For the purpose of this thesis, the presence of a bright band is not enough. Furthermore, it is required that precipitation is uniform over a wide range and a long period of time, i.e. that the horizontal variability of precipitation and therefore of radar reflectivity is as small as possible during the time period in question. However, slow variations of precipitation intensity as well as slow large-scale changes such as a descent of the melting layer or of the echo top should be accepted. Not accepted are embedded shower cells and convectivity.

These demands require a classification scheme independent of the vertical variability of radar reflectivity, thus based only on horizontal properties of radar quantities. The classification scheme described in two steps below meets all requirements stated above.
Fig. 3.5: Two PPIs and two RHIs recorded at two different times. The left two pictures were recorded when precipitation was at its maximum (4mm/h), the right two pictures when precipitation was weak (1mm/h). Marked with white dots in the PPIs are the locations of the optical spectrometer (left dot) and of the disdrometer/X-band radar (right dot). The direction of the RHI, intersecting the bottom station, is marked with a solid line. The vertical line in the RHIs indicates the position of the bottom station.
Means were calculated over:

a) 3 points, i.e. 0.6 km
b) 7 points, i.e. 1.4 km
c) 11 points, i.e. 2.2 km
d) 21 points, i.e. 4.2 km

Fig. 3.6: Radar reflectivities of the RHI from 21:43 have been smoothed horizontally over \( n \) points. Depicted are not the smoothed values of radar reflectivity but the standard deviations \( \sigma \). Note: the more points are used to smooth over, the more values are lost at the edges of the picture.

3.3.2 First step

To investigate horizontal properties of precipitation, RHIs are suited best. Looking at the RHI of 21:43 in Fig. 3.5 two facts are immediately perceptible: first, the stratiform character of the precipitation - horizontal structures and a bright band - second the uniformity over a range of 20 km.

To create the classification scheme, as a first step reflectivities are smoothed horizontally
3.3. A classification scheme

over \( n \) points. The emphasis is not on the mean values of reflectivity but on the standard deviations \( \sigma \). Fig. 3.6a shows the standard deviations resulting when smoothing radar reflectivities over \( n = 3 \) points. The majority of the values are less than 2 dB, more than half are less than 1 dB. This is not surprising since at this time the precipitation was very uniform and smoothing was done over just three points. It should be also noted that the error of radar reflectivities, as measured with the radar, are about 1 dB.

The horizontal resolution of the radar data is 200 m, i.e. smoothing over 3 points stands for smoothing over 600 m. This is not enough to show reliably horizontal uniformity. Figs. 3.6b–d show three more pictures with results of smoothing over 7, 11 and 21 points, respectively. A smoothing over 21 points, i.e. 4.2 km, seems to be reasonable and sufficient to show horizontal uniformity. A horizontal range of 4 km has already been used by Huggel et al. (1996) to identify stratiform precipitation. As can be seen, smoothing over more than three points does not increase the standard deviations. Picture d shows that still the majority of the points have standard deviations less than 2 dB. This is an evidence for the uniform structure of the precipitation.

Clearly visible in all four pictures in Fig. 3.6 is a black region at about 17 km. This is ground clutter (Mt. Albis), i.e. pixels without radar reflectivity. The C-band Doppler radar removes operationally all measurements with Doppler velocity equal to zero. Data points without precipitation cannot be distinguished from data points with zero Doppler velocity (be they ground clutter or not). If smoothing is done over \( n \) points, standard deviations will be set to 99 if one or more of the \( n \) points have no reflectivity information. This explains why information is lost at the edges of the picture and why the ground clutter of Mt. Albis is smeared out in picture d compared to picture a.

One glance at Fig. 3.6d is enough to realize that precipitation at 21:43 has been stratiform and extremely uniform. But, with this, no new insight is won since the same information is contained in the underlying RHI (Fig. 3.5 bottom left). The next step will show, how standard deviations can be used successfully to identify uniformity not only in space but also in time.

3.3.3 Second step

The case study of 17/18 February 1995 covers around 5.5 hours of precipitation with no significant changes of the precipitation structures. 172 RHIs have been taken and an inspection of all of them revealed the astonishing uniformity of the precipitation during the whole period.

In order to show this, standard deviations are calculated for all 172 RHIs as described in the first step. Smoothing is now done over 3 points. To summarize the results of all RHIs, standard deviations are sorted into the following five categories: 0–2 dB, 2–3 dB, 3–4 dB,
Fig. 3.7: Categories of standard deviations if smoothing has been performed over three points.

start: date 950217 time 19:21
end: date 950218 time 01:03
total number of RHIs: 172
means calc. over # of points: 3
3.3. A classification scheme

Fig. 3.8: Categories of standard deviations if smoothing has been performed over 21 points.

start: date 950217 time 19:21
end: date 950218 time 01:03

total number of RHIs: 172
means calc. over # of points: 21
4–10 dB and > 10 dB. Five pictures are generated, each belonging to one of these categories (see Fig. 3.7). These pictures are filled with information about the percentage of all RHIs with corresponding standard deviations. Thus, if picture 1 (0–2 dB) has at pixel $xy$ a value of 65 this signifies that 65 % of all 172 RHIs have at pixel $xy$ a standard deviation of less than 2 dB. In the fist step of the classification one single specific RHI had for most pixels a standard deviation of less than 2 dB. If all RHIs were similar to this one, one would expect that picture 1 (0–2 dB) in Fig. 3.7 would show overall quite high percentages.

The percentages in Fig. 3.7 are greyscale-coded with four levels: the darker the color the higher the percentage. As an example, picture 1 (0–2 dB) shows an almost uniform distribution of the color. At most pixels, 75 to 100% of the RHIs have a standard deviation of less than 2 dB. Picture 1 has a rather "dark appearance".

Picture 2 (2–3 dB) shows that at virtually all pixels less than 25% of the RHIs have standard deviations between 2 and 3 dB. Standard deviations of 3 to 10 dB are almost negligible, whereas no RHI has a pixel with standard deviation of more than 10 dB.

A sixth picture is added to these five categories of standard deviations. This picture contains information about pixels without reflectivity data. Again, there is at about 17 km the ground clutter of Mt. Albis where reflectivity data have been removed because of Doppler velocities equal to zero. This ground clutter and any other real ground clutter has a percentage of occurrence of 100 and is therefore drawn black in picture 6. But there are other pixels in picture 6 where the radar did not measure reflectivities for some RHIs. There is, for example, in the upper right corner a section where up to 50% of the RHIs had no reflectivities. This has been caused by the descending of the echo top during the precipitation event. These pixels are real measurements without radar echo. Below 2 km of height, there are also pixels without reflectivities. There is, especially, a thin layer around 1.2 km. Reflectivity data for these pixels have been removed falsely. For some time, the wind was blowing at this height perpendicular to the azimuth of the RHIs, thus measured Doppler velocities were near zero and therefore reflectivity data were removed.

Smoothing over 7, 11 and 21 points have also been done and the main findings are the same as described above for smoothing over 3 points. Fig. 3.8 shows the results for smoothing over 21 points. The percentages of occurrence of 0–2 dB have decreased somewhat but still the majority of the RHIs have standard deviations for almost all pixels below 2 dB. It is also an evidence of the uniformity of the precipitation that standard deviations of more than 3 dB decrease between smoothing over 3 and 21 points. Smoothing over 21 points is more reasonable to show horizontal uniformity than smoothing over just 3 points. Although for this case there is not much difference when smoothing is done over 3 or 21 points, in non-uniform precipitation differences become apparent. In appendix A two more cases, one stratiform with embedded showers and one thunderstorm are shown to illustrate this.

Therefore, Fig. 3.8 is proposed to serve as a classification scheme for precipitation and
especially as a tool to detect stratiform und uniform (in space and time) precipitation. It is proposed that precipitation periods where the "darkest appearance" is shown by picture 1 (0–2 dB) fulfill the requirements stated in Sect. 3.3.1.
3.4 Overview of the data

In this section data obtained with the different instruments will be illustrated and discussed shortly. Since the data of the C-band radar already have been discussed, this section is devoted to the instruments at the bottom and top stations at Mt. Üetli.

3.4.1 Bottom station – X-band Doppler radar

The vertically pointing X-band radar measured profiles of radar reflectivity and Doppler velocity from 50 m to 14.25 km AGL with a time-resolution of 30 s. A 3-point smoothing has been applied to these profiles. Height-time-diagrams (HTI) of radar reflectivity and Doppler velocity of the X-band radar are presented in Fig. 3.9. A well developed bright band with maximum reflectivity of more than 35 dBZ and with a vertical extension between 300 m and 550 m is visible. Fig. 3.9 shows nicely how the bright band descended slowly and steadily during the 5.5 hours with a rate of about 80 m/h and how the optical spectrometer on top of Mt. Üetli (its position is marked with a white line) first measured rain, then melting snow and at last dry snow.

Doppler velocities above the melting layer up to 3.5 km are 0 to 2 m/s. First, this shows that no upward winds with velocities exceeding a few 10 cm/s are present, second, this indicates that the snow particles are at most moderately rimed. Particles with higher riming have fall velocities exceeding 2 m/s (Mosimann, 1994a). However, around 23:30 Doppler fall velocities right above the melting layer are less than 1 m/s indicating that upward winds have sped up for a short time.

3.4.2 Bottom station – disdrometer

Disdrometer data were recorded with an integration time of 60 s. Depletion of the smaller drops due to the dead time of the disdrometer was calculated and then corrected (Sauvageot and Lacaux, 1995). Drop size spectra are assumed to be exponential. Thus, the drop size distribution $N(D)$ can be represented by

$$N(D) = N_0 e^{-\Lambda D}, \quad (3.1)$$

according to Marshall and Palmer (1948), with the two constants $N_0$ (m$^{-3}$mm$^{-1}$) and $\Lambda$ (mm$^{-1}$). $N_0$ and $\Lambda$ are calculated from the measured drop size spectra. Both parameters are obtained from a transformation using the liquid water content $W$ and the radar reflectivity factor $Z$, calculated from the measured raindrop size distributions:

$$W = \frac{\pi}{6} \int_0^\infty N(D)D^3dD = \frac{\pi}{6} \sum_{i=1}^{36} N_i(D_i)D_i^3 \Delta D_i \quad (3.2)$$
3.4. Overview of the data

Fig. 3.9: Height-time diagrams of radar reflectivity and Doppler velocity of the vertically pointing X-band Doppler radar. The height of the optical spectrometer at the top station is indicated with the horizontal white line. The gaps without data are due to the change of tapes.
This transformation ensures that, since the exponential fit does not represent the measured distribution exactly, at least radar reflectivity and liquid water content, the two most important quantities of precipitation, are conserved. Substituting Eq. 3.1 in the Eqs. 3.2 and 3.3, the parameters $N_0$ and $\Lambda$ can be calculated as follows:

$$N_0 = \frac{1}{\pi} \left(\frac{6!}{\pi}\right)^{\frac{3}{2}} \left(\frac{W}{Z}\right)^{\frac{3}{2}} \cdot W$$

$$\Lambda = \frac{6!}{\pi} \left(\frac{W}{Z}\right)^{\frac{1}{2}}.$$

The quantity $W$ is in mm$^3$/m$^3$ and $Z$ in mm$^6$/m$^3$. The parameters $N_0$ and $\Lambda$ are quantities, which agree exactly with the geometrical intercept and slope parameter only if the measured raindrop size distribution is of an exact exponential shape. In all other cases, however, the parameters $N_0$ and $\Lambda$ are very useful quantities to get an idea about the type of the spectra considered: Small drop spectra show large values of $N_0$ and $\Lambda$, whereas large drop spectra have small values of $N_0$ and $\Lambda$ when compared with the "classical" values from Marshall and Palmer (1948).

Fig. 3.10 gives an overview of the parameters of the drop size distribution measured by the disdrometer as well as the rainrate and the rainrate measured with a tipping bucket. $N_0$ and
A reveal that the drop size distribution was somewhat variable throughout the experiment. While drop spectra with large drops were measured during the first 3.5 hours, drop sizes decreased during the last 2 hours. Rainrate was also changing, first increasing from 1 to 4 mm/h, then decreasing again to 1 mm/h with one short period of rainrate below 0.5 mm/h. Total rainrate was 9.6 mm.

3.4.3 Top station – optical spectrometer

Particle size and velocity distributions have been calculated from the data measured with the optical spectrometer with the restrictions discussed in Sect. 2.2.1. Wind was negligible, thus no errors due to wind effects are considered.

Particle size distributions

Particle size distributions are depicted in Fig. 3.11. Average distributions have been calculated for 15-minute intervals. Three intervals are missing (20:30, 20:45, 21:15) due to a system breakdown. Maximum hydrometeor sizes depicted are limited to 15 mm in order to enhance the clarity at small particle sizes. However, between 21:45 and 22:45 large snowflakes with sizes well over 20 mm have been sampled. Thus, spectra for these intervals are truncated in Fig. 3.11.

The transition from rain to snow is monitored step by step in Fig. 3.11. First, rain is measured until 21:00 and maximum drop sizes do not exceed 5 mm. All distributions have a distinct exponential shape. Suddenly, at 21:30, large particles appear. The size distribution is no more exponential. Yet, it seems that hydrometeors up to 3 mm have an exponential distribution. These hydrometeors are completely melted, i.e. raindrops, and their distribution is similar to the distributions measured before 21:00. The larger hydrometeors are not yet melted. Some will be in an advanced state of melting, but the largest ones have probably just started to melt. The same as for the interval at 21:30 holds for the intervals until 23:00. The part with completely melted drops, however, becomes less visible as time moves on and reflects the fact that the largest melted drop becomes smaller and smaller as the melting layer descends. After 23:00 no evidence can be found that the spectra might be composed of two different kinds of particles, melted and partly melted, despite the fact that the optical spectrometer is still well below the top of the melting layer (see radar reflectivity, Fig. 3.9).

Particle velocity distribution

Again, the melting of particles is monitored by the fall velocity distributions in Fig. 3.12 step by step. At the beginning, particle fall velocities are equal to the fall velocity of raindrops given by Atlas et al. (1973) after Gunn and Kinzer (1949). Fall velocities of large raindrops
Fig. 3.11: Hydrometeor size distribution measured with the optical spectrometer on top of Mt. Üetli. Spectra have been averaged over 15 minutes.

are somewhat larger than Gunn and Kinzer predict. This is a consequence of the evaluation software of the optical spectrometer which neglects the fact that large raindrops are oblate and hence overestimates the fall velocities of large raindrops.

Suddenly, large hydrometeors appeare, the fall velocities of which differ considerably from the Gunn and Kinzer curve. At 21:30, two regimes are clearly visible. Particles with sizes up to approx. 1.5 mm have fall velocities equal to the velocity of raindrops. These particles are completely melted. Particles larger than 1.5 mm have, with growing size decreasing fall
Fig. 3.12: Hydrometeor velocity distributions measured with the optical spectrometer on top of Mt. Üetli. Spectra have been averaged over 15 minutes. As a comparison, the fall velocity of raindrops by Atlas et al. (1973) and the fall velocity of snowflakes by Zikmunda (1972) and Locatelli and Hobbs (1974) are printed in each picture.

velocities. The largest particles have a fall velocity of below 2 m/s. The decreasing fall velocity with increasing size illustrates the different stage of melting. While particles around 2 to 3 mm are almost completely melted, large particles with sizes over 10 mm have just begun to melt, if at all. The next few intervals show a similar behaviour with a clearly visible upper limit for the size of the completely melted drops. This limit decreases from 1.5 mm
at 21:30 to about 0.5 mm at 22:15.

No upper limit for the size of the completely melted drops can be found in the spectra after 22:15. The fall velocities of the particles are similar to the fall velocity given by Locatelli and Hobbs (1974, dotted line) for aggregates of side planes and by Zikmunda (1972, dashed line) for general aggregates. The largest hydrometeors seem to differ considerably from the curve given by Zikmunda (1972), having higher fall velocities. One possible explanation is that the snowflakes are oblate and thus their fall velocity is overestimated by the evaluation software (see Sect. 2.2.1).

3.4.4 Top station – Formvar replica

To determine the snow particle types and the amount of snow rime, snow particles have been collected by the Formvar replication method (Schaefer, 1956) at the top station. Formvar coated slides were exposed to the precipitation for the short period of a few seconds to gather ice crystals. Since precipitation at the top station gradually changed from rain to snow, this was only possible during the last two hours. The amount of snow rime was determined according to the classification of Mosimann et al. (1994b). This classification is defined as a six-step scale of snow rime, running from 0 (unrimed) to 4 (heavily rimed) and 5 (graupel).

The degree of snow rime during the last two hours was evaluated to be between unrimed and lightly rimed. As mentioned above, precipitation was changing from rain to snow. As soon as possible, Formvar replica were taken, even if the conditions were not ideal. Usually, Formvar replicas are best if taken at subzero temperatures, preferably below -1°C. This was not the case in this study. At the top station, the temperatures at the end of the experiment were somewhat below 0°C, at the time Formvar replication started, maybe at approx. 1°C. This means not only that ice particles might have had a thin coat of water covering their surface prior to falling onto the slides, but also that the high temperature of the Formvar solution could have added to the melting. It is possible that these effects had an influence on the estimation of the rime degree in the sense of underestimating the rime.

A relationship between the rime degree of snow particles above the melting layer and the parameters of the drop size distribution measured at the ground has been found by Barthazy et al. (submitted). High rime degrees have been observed to correlate with large values of $N_0$ and $\Lambda$, i.e. a small drop size distribution, while unrimed snowflakes seem to yield a raindrop spectrum with large drops, i.e. small values of $N_0$ and $\Lambda$. Considering the time series of $N_0$ and $\Lambda$ of this case study shown in Fig. 3.10, ice particles had probably lower rime degrees during the first 3.5 hours of the experiment than during the last 2 hours. Thus, rime of snow particles throughout the whole experiment is assumed to be at most moderate.
3.4. Overview of the data

The predominant measured types of snow crystals were irregular crystals, side planes, columns and platelike snow crystals. No significant number of needles or dendrites were found. Snow particle type and occurrence in the alpine region has been investigated by Mosimann (1994a). A total of more than 700 Formvar replicas with about 40,000 ice crystals have been taken during several precipitation events. More than 55% of the crystals sampled belong to irregular or platelike types. Needles, columns and bullets amount for another 15% and 16% of dendrites are reported. Although dendrites were observed occasionally during the case study, their portion is less than 16%. Nevertheless, the mixture of crystal types observed are not unusual.
3.5 The position parameter H

As the hydrometeor size and velocity distributions were discussed in the previous section, the observed features could have been compared with the radar profiles at the same time in order to know the position of the optical spectrometer within the melting layer at the time in question. One possibility to give the position of the optical spectrometer within the melting layer is to give the vertical distance to the top of the bright band. The top of the bright band is observed to correspond with the 0°C isotherm (Fabry and Zawadzki, 1995), i.e. to mark the beginning of the melting of ice particles. Thus, the top of the bright band seems to be an important reference point with the physical interpretation as the height where melting starts. Therefore, a quantity \( \Delta h \), defined as the vertical distance to the top of the bright band, could be used to specify the position of the optical spectrometer within the melting layer. But the vertical extent of the bright band strongly depends on the rainrate. Without knowing the rainrate, a position of, e. g., \( \Delta h = 400 \) m below the top of the bright band cannot be identified as being within or below the melting layer. This is a disadvantage of the quantity \( \Delta h \).

However, a radar profile yields more information about the melting layer than just the top of the bright band. In the following, a position parameter \( H \) will be defined which will be used throughout the rest of this thesis to refer to as a position within the bright band.

3.5.1 The definition of the position parameter \( H \)

![Fig. 3.13: Radar reflectivity profile with a well developed bright band. The three horizontal lines indicate three possible positions of the optical spectrometer with the corresponding values of \( H \): \( H = 0 \) maximum of radar reflectivity, \( H = 1 \) top of the bright band, \( H = -1 \) bottom of the bright band (see text for further explanation).](image)

Every radar profile with a pronounced bright band and a sufficient vertical resolution shows three characteristic features: the top, the bottom and the maximum of the bright band. The top of the bright band marks, as mentioned before, the beginning of the melting of ice particles. The bottom of the bright band is generally believed to mark the end of melting, i.e. to be at the height where all the ice particles have melted to raindrops. However, the maximum of the bright band is not so obviously connected to melting alone, but is a consequence of the sum of several microphysical properties such as the dielectric constant.
3.5. The position parameter $H$

of the melting flake, the distribution of the melted water within the flake, the fall velocity of flakes and hence of the concentration of the flakes within space and the orientation of nonspheric snowflakes, just to specify the more important properties. Nevertheless, the maximum of radar reflectivity is the most prominent feature of a stratiform radar profile and will be included in the definition of the position parameter $H$.

First, the top of the bright band is determined according to Fabry and Zawadzki (1995). The new position parameter $H$ is set to one at the top of the bright band and to zero at the maximum of radar reflectivity. Fig. 3.13 shows a profile of radar reflectivity. The solid horizontal line is the position of the maximum radar reflectivity ($H = 0$) and the dashed line indicates the top of the bright band ($H = 1$). With this, the part above the maximum of radar reflectivity has positive values of $H$. Since radar profiles are usually not symmetric, in a second step $H$ is set to -1 at the bottom of the bright band. With this, the part below the maximum of radar reflectivity has negative values of $H$.

This parametrization has many advantages. First of all, measurements made in different precipitation events with the $0^\circ\text{C}$ isotherm at different heights can be compared easily. A position parameter of, e.g., $H = 0.5$ indicates that measurements were made in the middle of the upper part of the melting layer, independent of its actual height above ground level. Another advantage is the elimination of the effect of different rainrates. The vertical extension of the melting layer and hence of the bright band depends on the rainrate. With the conception of the position parameter $H$, cases with different rainrates may easily be compared.

It is proposed that the position parameter $H$ is a much better reference to describe microphysical properties within the melting layer than the vertical distance to the top of the bright band. Microphysical observations at the same position $H$ of different cases are assumed to be more consistent than at the same vertical distance below the top of the bright band. Two examples, proving the validity of this assumption, can be found in the sections 5.1.2 and 5.2. Thus, from now on, the position parameter $H$ is used to discuss microphysical observations within the melting layer.

Usually, the independent quantity of a relation is depicted on the horizontal axis while for the dependent quantity the vertical axis is used. However, in the discussion of microphysical observations the independent quantity $H$ will be depicted on the vertical axis. This seems to be natural since it stands for a vertical quantity.

Some applications of the position parameter $H$ will be shown in the next sections.
3.6 Discussion of the case study

3.6.1 The largest particle measured with the optical spectrometer

In Sect. 3.4.3 it has been shown that particle sizes, measured with the optical spectrometer, change rapidly during the measurements. First, particles are small but grow quickly as the optical spectrometer measures around the maximum of radar reflectivity. Sizes decrease again as the optical spectrometer measures within the upper part of the melting layer.

The largest particle measured in each one-minute-interval may be depicted in dependence of the previously defined position parameter $H$ as Fig. 3.14 shows. Within snow, i.e. if the optical spectrometer measures above the bright band, maximum particle sizes of about 7 mm are measured. No change of this value is observed within the uppermost part of the bright band. But maximum particle sizes increase rapidly as the optical spectrometer measures somewhat above the maximum of radar reflectivity. The largest particles throughout the whole experiment, which are well over 20 mm, are observed slightly below the maximum of radar reflectivity. Within the lower part of the bright band maximum particle sizes decrease again and below the bright band particles have maximum sizes of 2 to 3 mm.

The increase of the maximum sizes of particles around $H = 0$ is assumed to be mostly due
3.6. Discussion of the case study

to aggregation. However, the rainrate is not constant during the measurements but was maximal when the optical spectrometer was measuring at the height of maximum radar reflectivity. The large particles observed at the position $H = -0.1$ could be a result not just of aggregation but also of the increased rainrate. To estimate the quantitative influence of an increased rainrate on hydrometeor sizes, the following consideration is made:

- At 21:55, which corresponds to a position $H = -0.14$ the largest snowflakes (22 to 28 mm) are observed. At this time the rainrate is between 3 and 4 mm/h. Later, at 22:50 the rainrate is again between 3 and 4 mm/h. This time corresponds to the position of $H = 0.20$ and maximum snowflake size is 12 mm. Hence, the rainrate does not seem to be responsible for the large sizes of snowflakes observed around $H = -0.1$.

3.6.2 The largest melted particle

In the section with the discussion of the fall velocities measured with the optical spectrometer (Sect. 3.4.3), the largest melted particle was identified for some time intervals. This largest melted particle can now be depicted in dependance of the position parameter $H$ as shown in Fig. 3.15. The identification was done visually, simply by comparing the measured fall velocities with the fall velocity of raindrops given by Atlas et al. (1973). From Fig. 3.15 it can be seen that within the upper part of the melting layer (22:20 to 01:00) almost no particles are completely melted. The largest melted particle at the maximum of radar reflectivity ($H = 0$, 22:20) has a diameter of about only 0.5 mm.

Fig. 3.15: The largest melted particle in dependence of the position parameter $H$. The smallest detectable limit for the largest melted particle is around 0.2 mm.
3.6.3 The particle number flux

To compare the measurements made with the optical spectrometer with the data obtained with the disdrometer, a quantity called the number flux is introduced. The number flux is defined as

\[
\text{number flux} = NF = \sum_i \frac{n_i}{A_i},
\]

where \(n_i\) is the number of particles sampled during one minute in the size class \(i\) and \(A_i\) is the size of the corresponding measuring area. The number flux can be defined for the optical spectrometer as well as for the disdrometer. For the disdrometer, \(A_i\) is assumed to have the same value for all size classes (50.3 cm², see Sect. 2.4), for the optical spectrometer \(A_i\) depends on the mean diameter \(d_i\) of class \(i\) (see Table 2.2). A consequence of this definition is that the number flux is independent of the measured fall velocity of the hydrometeors. Thus, the limited abilities of the optical spectrometer to measure exact fall velocities does not affect this quantity. In Fig. 3.16, the number flux at the disdrometer and at the optical spectrometer is shown in dependence of the position parameter \(H\). Here it has to be noted that the position parameter \(H\) is a reference to the height only of the optical spectrometer within the melting layer. It does not contain any information about the position of the disdrometer. The disdrometer stands throughout the whole experiment at the bottom station and measures rain from the begin to the end. If the NF at one of the instruments is given e.g. at \(H = -0.5\) this means that the optical spectrometer measures melting particles at \(H = -0.5\) within the melting layer while the disdrometer measures rain at some distance below the melting layer. However, the NF at the disdrometer can also be depicted in dependence of the position parameter \(H\), simply by using the same transformation from time to \(H\) as was used to calculate the position of the optical spectrometer.

When comparing the NF at the two instruments, the smallest particle detectable with the two instruments has to be considered. The smallest size class of the optical spectrometer has a diameter of 0.146 mm and of the disdrometer a diameter of 0.36 mm. All raindrops with sizes between 0.146 and 0.36 mm can be detected with the optical spectrometer but not with the disdrometer. In order to be able to compare the particle fluxes correctly, the first two size classes are omitted in the summation of the particle flux at the optical spectrometer. Thus, a threshold is set for the optical spectrometer with the smallest particles counted being those with a diameter of 0.438 mm.

However, if the optical spectrometer measures dry snow, an ice particle of this size may well be much smaller when melted (and falling onto the disdrometer) and thus remain undetected by the disdrometer. That is, a fixed threshold for the optical spectrometer is not appropriate for all kind of hydrometeors. If dry snow is measured, another threshold has to be used. To compare the number flux at both instruments when the optical spectrometer measures snow, a new threshold is introduced by calculating the size of a snowflake that yields a raindrop with a diameter of 0.36 mm (instrumental threshold of the disdrometer).
Locatelli and Hobbs (1974) give mass-size relationships for different types of ice particles and snowflakes. Aggregates of unrimed side planes seem to correspond best to the observed type of snowflakes, thus this relationship is used:

$$ M = 0.04 \cdot D^{1.4} \quad 0.5 < D < 4.0 \text{ mm} \quad (3.4) $$

where $M$ is the mass of the snowflake in mg and $D$ the diameter in mm. Furthermore, it has to be considered that particles have a riming degree $Rm$ between 0 and 2. Mosimann et al. (1994b) give the mass $M_r$ of rimed snow particles in dependence of a rimed mass...
fraction \( f \)

\[
M_r = \frac{M}{1 - f} \quad f = \frac{0.017(3.3R_{\text{rim}} - 1)}{1 + 0.017(3.3R_{\text{rim}} - 1)}
\] (3.5)

where \( M \) is the unrimed mass of the snow particle and \( R_{\text{rim}} \) the riming degree. Combining the Eqs. 3.4 and 3.5, the size \( D_{\text{min}} \) of the smallest snowflake detectable with the disdrometer below the melting layer can be calculated:

\[
D_{\text{min}} = \left( \frac{\pi}{6} D_{\text{melt}}^3 \rho_w (1 + 0.017(3.3R_{\text{rim}} - 1)) \right)^{1/4}
\] (3.6)

where \( D_{\text{melt}} \) is the melted diameter of the snowflake, i.e. \( D_{\text{melt}} = 0.36 \) mm. \( D_{\text{min}} \) has the size:

\[
0.70 < D_{\text{min}} < 0.78 \text{ mm} \quad \text{(for } 0 \geq R_{\text{rim}} \geq 2)\]

In a second attempt, particle number fluxes are calculated at the optical spectrometer with this higher threshold: not just the two smallest size classes are omitted in the summation but the four smallest ones. Thus, the first size class taken into the sum is class five with a diameter of 0.73 mm. The right picture in Fig. 3.16 shows mainly the same as the left one but the number fluxes at the optical spectrometer are now lower and only reasonable for \( H \approx 1 \).

### 3.6.4 The number flux ratio

In order to compare the number fluxes at the optical spectrometer and the disdrometer, a second quantity is introduced which is called the number flux ratio (NFR). The number flux ratio is defined as

\[
\text{number flux ratio} = \text{NFR} = \frac{\text{NF}_{\text{optical spectrometer}}}{\text{NF}_{\text{disdrometer}}}.
\]

The number flux ratio is equal to one if both instruments measure the same number of particles and smaller than one if the optical instrument measures less particles than the disdrometer.

Calculating the NFR, the problem with different thresholds for the optical spectrometer due to different types of hydrometeors has to be solved. A completely satisfying solution is not possible. A gradual change between the threshold for raindrops and for snowflakes would, theoretically, solve this problem but is not desirable since it would complicate the calculations enormously and clarity would be lost. Thus, in order to keep the continuity, the threshold for raindrops is used when calculating the NFR for all kind of hydrometeors. With this, an error is introduced to the NFR when dry snow is measured. However, as soon as melting starts, the smallest particles are the first to melt. As Fig. 3.15 shows, the largest melted particle at \( H = 0 \) has a diameter of about 0.5 mm. It seems to be reasonable
3.6. Discussion of the case study

to assume that particles of the size class three (0.4 mm, the smallest taken into the sum) are completely melted up to values of $H$ around 0.3, i.e. the NFR is correct for values of $H < 0.3$. For $H > 0.3$, the number of ice particles taken into the sum that have melted diameters below the instrumental threshold of the disdrometer gradually increase and thus the number flux ratio is affected with errors. This has to be considered when evaluating the NFR.

![Diagram showing the number flux ratio in dependence of the position parameter $H$.](image)

The number flux ratio in dependence of the position parameter $H$ of this case study is shown in Fig. 3.17. The features and the implications of this relationship will be discussed now step by step.

a) At $H \approx 1$, the optical spectrometer measured dry snow at the top of the melting layer while the disdrometer measured rain below the melting layer. Here it has to be considered that a part of the number flux at the optical spectrometer is below the instrumental threshold of the disdrometer and the NFR is not correct. But the right picture in Fig. 3.16 shows the corrected number flux at the optical spectrometer when dry snow is measured and it can be seen that the number flux is equal for both instruments. Thus, the NFR is in reality equal to one.

b) From $H = 1.0$ to $H = -0.10$, the NFR is continuously decreasing. This decrease of the NFR, caused by a decrease of the NF at the optical spectrometer, could be explained
by aggregation of hydrometeors dominating over breakup. Aggregation would also lead to the large particle sizes measured at $H = -0.15$.

c) Around the height of maximum radar reflectivity, i.e. at $0.2 > H > -0.15$ the optical spectrometer measured raindrops and partly melted particles (discussed in Sect. 3.4.3) while the disdrometer continuously measured rain. The number flux ratio is now around 0.4, indicating that the particle flux at the optical spectrometer is by a factor of 2 to 3 smaller than the particle flux at the disdrometer.

d) Between $H = -0.10$ and $H = -0.8$ the NFR is increasing again. This increase of the NFR is accompanied by a rapid decrease of maximum particle sizes (see Fig. 3.14) and could be explained by breakup of hydrometeors. The decrease of maximum particle sizes is most probably not only due to the collapse of melting particles but also to breakup of hydrometeors.

e) At $H < -0.8$, NFR is equal to one, i.e. the number flux at both instruments are about the same. This is not surprising since both instruments measure rain and the vertical distance between the two instruments is only 450 m.
3.7 Conclusions of this case study

The comparison of the particle fluxes measured with two instruments, one sampling rain and one sampling all kind of hydrometeors within the melting layer lead to some tempting assumptions:

1. At the end of the experiment, the optical spectrometer measured right at the top of the bright band. The vertical extension of the bright band equalled at this time the vertical distance between the top and the bottom station. Thus, the disdrometer sampled raindrops just at the bottom of the bright band. As it was already noted, the number flux at both instruments is equal at \( H \sim 1 \). This seems to imply that, whatever is happening within the melting layer, one snowflake yields, on the average, one raindrop. However, this does not say that one individual snowflake melts to exactly one individual raindrop, but the total number of snowflakes is equal to the total number of raindrops.

Experimental verification of this assumption is scarce and rudimentary (e.g. Ohtake, 1969) and the assumption itself is not believed by common consent. While some researchers accept this assumption as quite reasonable (Drummond et al., 1996, Ekpenyong and Srivastava, 1970) others do believe that processes within the melting layer alter the size distribution too much to sustain this assumption (Gunn and Marshall, 1958). After all, most researcher rely on considerations based upon indirect methods. No extensive direct observations are available up to now.

2. The decrease of the NFR, caused by the decrease of the number flux at the optical spectrometer within the upper part of the melting layer seems to be an effect of aggregation. This assumption is strengthened by the fact that particle sizes increase rapidly within the upper part of the melting layer. It seems to be obvious that aggregation is active and quite effective within the melting layer. This is not a surprise since most researchers, above all those working with radar, believe that aggregation is active within the melting layer, the question is only to what extent.

3. The increase of the NFR, caused by the increase of the number flux at the optical spectrometer within the lower part of the melting layer seems to be an effect of breakup. Aggregations still could be active, but seems to be dominated by breakup.

In the literature, much less attention is given to the breakup of hydrometeors within the melting layer than to aggregation. Few authors include the concept of breakup into their theories about the melting layer (e.g. Fabry and Zawadzki, 1995). Again, no direct measurements exist dealing with breakup.

The thorough evaluation of this case study yields some promising results of the experimental detectability of aggregation and breakup within the melting layer. To test these results, more
Experimental data are needed. In the next chapter, three additional cases will be introduced which are used in Chapter 5 to verify the results of this chapter.
Chapter 4

Additional cases

After the encouraging results obtained with the measurements at Mt. Úetli, a winter campaign was planned to measure similar cases of stratiform precipitation. Since the vertical distance between the bottom and the top station at Mt. Úetli was barely enough (450 m) for observations within the melting layer, the measuring site was dislocated to Mt. Rigi, a precursor of the Alps in central Switzerland. Here, three stations were set up along a steep slope with more or less the same instrumentation as on Mt. Úetli. The main difference was the use of the 2D-Video-Disdrometer, a second optical spectrometer, which was installed on the third, the top station. This instrument was used during five months of the winter 96/97 as a loan from Joanneum Research in Graz, Austria.

Precipitation in the winter season (November to March) of 96/97 was scarce, with only 10% of the average normal in January and the beginning of February. However, the situation improved mid February and rainrates reached almost normal values in March. The requirements of a precipitation event to be included into this study were quite severe, thus, not more than three cases could be found. This chapter will shortly illustrate these three cases in a similar manner as the case of Mt. Úetli has been illustrated in the previous chapter. All cases together will be discussed in Chapter 5.

4.1 Experimental setup

Fig. 4.1 shows the setup along the northwestern slope of Mt. Rigi. The bottom station is located at Greppen (448 m asl), the middle station on the Seebodenalp (1031 m asl) and the top station at Rigi Staffel (1603 m asl). The horizontal distance between the bottom and the middle station is 3.0 km and between the bottom and the top station 3.5 km.

The bottom station was equipped, as in the case of Mt. Úetli, with the mobile X-band Doppler radar, the disdrometer and meteorological instruments. The middle station housed the optical spectrometer and Formvar replication facilities. Meteorological parameters on
Fig. 4.1: Side view and map of Mt. Rigi with the three stations along its northwestern slope. Top picture reproduced with the permission of the Rigi-Bahnen, bottom picture reproduced with the permission of the Bundesamt für Landestopographie, 28 April 1998
Seebodenalp were measured by an operational NABEL measuring unit (NAtionales BEobachtungsnetz für Luftfremdstoffe) operated by the BUWAL (Bundesamt für Umwelt, WALd und Landschaft), and were kindly made accessible by the BUWAL. At the top station, the 2D-Video-Disdrometer was placed as well as meteorological instruments and Formvar replication facilities. The C-band Doppler radar at the ETH in Zürich in a distance of 40 km recorded PPIs and RHIs. SMI composit radar data were also used as in the case of Mt. Üetli.

4.2 Selection of good cases

Cases similar to the case at Mt. Üetli had to be found, thus, a case to be included into this study had to fulfill the following criteria:

1. The melting layer had to be either descending or ascending,

2. the melting layer had to pass either the top station with the 2D-Video-Disdrometer or the middle station with the optical spectrometer,

3. precipitation had to be very uniform to be able to compare data measured at the different stations that were a few kilometers apart,

4. the time series of particle fluxes within the melting layer and of rain had to show the same features, i.e. maxima and minima had to be coincident more or less.

The first criterion is necessary to obtain a profile of microphysical parameters within the melting layer with instruments having a fixed position. This limits the potentially interesting meteorological conditions to a case with a front passing the measuring site. The disadvantage of such a case is that the passage of a front is most often accompanied by more or less strong winds limiting the uniformity of precipitation. The second criterion is obviously necessary. The third and the fourth criterion were the critical points in selecting or rejecting a case. Uniformity of precipitation was judged by evaluating the RHIs recorded with the C-band radar but the discrimination between a good and a bad case was mainly done with the use of criterion 4.

These criteria were, as already mentioned, very limiting. The winter campaign yielded totally 110 hours of data of precipitation. Out of these 110 hours, only 3 hours could be used. Tab. 4.1 summarizes some of the main features of these three cases.
Table 4.1: Some parameters of the three selected cases. The predominant crystal type names the clearly recognizable crystal type most often found. However, for all cases, most crystals cannot be classified and belong thus to the irregular type.

<table>
<thead>
<tr>
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<tbody>
<tr>
<td>Duration</td>
<td>13:00–13:50</td>
<td>10:00–11:20</td>
<td>3:00–3:50</td>
</tr>
<tr>
<td>$\bar{R}$ (mm/h)</td>
<td>0.93</td>
<td>2.74</td>
<td>5.06</td>
</tr>
<tr>
<td>$R_{\text{tot}}$ (mm)</td>
<td>0.79</td>
<td>3.70</td>
<td>1.69</td>
</tr>
<tr>
<td>Melting Layer</td>
<td>Rigi Staffel</td>
<td>Seeboden</td>
<td>Seeboden</td>
</tr>
<tr>
<td>Rain</td>
<td>Seeboden</td>
<td>Greppen</td>
<td>Greppen</td>
</tr>
<tr>
<td>Predominant crystal type</td>
<td>Dendrites, plates</td>
<td>Needles, solid dendrites</td>
<td>Delicate dendrites</td>
</tr>
<tr>
<td>Average rime degree</td>
<td>0 to 1</td>
<td>2 to 3</td>
<td>1 to 2</td>
</tr>
</tbody>
</table>

4.3 Synoptic situation

The synoptic situations of these three cases were somewhat different from the situation at Mt. Üetli. For two of the cases, also cold fronts were responsible for precipitation and a descending of the melting layer, but the changes of meteorological conditions took place much faster. The rate of descent of the melting layer was 1.4 m/h for the case at Mt. Üetli whereas the new cases show rates of 4.4 m/h and 8.9 m/h. One case was not initiated by a cold front, thus the descent of the melting layer was marginal and data could be gained only of a confined part of the melting layer.
4.4 Overview of the three cases

A classification of the cases was done according to the scheme established in Sect. 3.3. The only difference to the classification of the case at Mt. Üetli is the distance between the radar and the experimental site. The radar beam is wider near Mt. Rigi than near Mt. Üetli. Thus, standard deviations are generally higher for the classification of cases at Mt. Rigi. The results for all three cases are shown in the three Figs. 4.2, 4.4, 4.6. In addition to this classification, microphysical parameters of the three cases are summarized and displayed in the Figs. 4.3, 4.5, 4.7. The summary consists of the height-time diagramm of radar reflectivity and time series of rainrate, temperature and number fluxes.

Number fluxes are plotted for raindrops and for hydrometeors within the melting layer. The appropriate station at which these data were measured can be seen in Tab. 4.1. The time series of rainrate give the rainrate measured at the same station as the NF of raindrops.

14 February 1997

Precipitation was initiated by a weak warm front passing Mt. Rigi on 14 February 1997. Temperatures were rising from midnight until noon and remained afterwards constant for the rest of the day. Precipitation started at around 10:00 with maximum rainrate at around 13:00. However, rainrate was rather low for the whole event. Two time intervals with a well developed bright band have been selected for this case. The melting layer was for these two intervals approx. at the height of the top station on Mt. Rigi and rain was measured at the middle station. The time series of NF at both, the top and the middle station show similar features.

15 February 1997

Following the warm front on 14 February 1997 a cold front passed Mt. Rigi between 10:00 and 11:00 on 15 February 1997. Precipitation started at around 9:30 and lasted at least until 17:00. The melting layer descended from above 1000 m ASL to the ground (450 m ASL) within the first two hours. This case was extremely uniform, even better than the case at Mt. Üetli. However, time series of the NF at the middle and the bottom station show less correspondence than on 14 February 1997.

19 March 1997

On 19 March 1997 precipitation was again initiated by a cold front passing Mt. Rigi early in the morning. Precipitation started at 02:00 and lasted for 2.5 hours. The rainrate was rather high during the selected interval. The time series of the NF at the bottom and the
Fig. 4.2: Classification-chart for 14 February 1997. Ground clutter around 40 km is Mt. Rigi itself. Precipitation was very weak during this case. However, while reflectivity in regions below 0.5 km is variable due to the restricted visibility of the radar, precipitation seems to be quite uniform between 0.5 and 2.5 km ahead of Mt. Rigi.
Fig. 4.3: Summary of microphysical properties for 14 February 1997. The two shaded regions indicate the periods of which data are used for this study. Temperature measurements at the middle station were not reasonable for this time period and are thus left away.
start: date 970215 time 10:02
end: date 970215 time 11:22
total number of RHIs: 9
means calc. over # of points: 21

Fig. 4.4: Classification-chart for 15 February 1997. Some variation is observed around Mt. Rigi and at lower elevations, but ahead of Mt. Rigi virtually no variation is observed.
Fig. 4.5: Summary of microphysical properties for 15 February 1997. The shaded region indicates the period of which data are used for this study.
start: date 970319 time 03:02
end: date 970319 time 03:52
total number of RHIs: 6
means calc. over # of points: 21

Fig. 4.6: Classification-chart for 19 March 1997.
4.4. Overview of the three cases

Fig. 4.7: Summary of microphysical properties for 19 March 1997. The shaded region indicates the period of which data are used for this study.
middle station are again in good agreement.

4.4.1 Comparing number flux densities

To compare the number fluxes at two different stations, a time lag of the data has to be considered. Such a time lag can be a result of the fall time of hydrometeors between the two stations and of the horizontal advection of precipitation structures.

The sources and effects of a possible time lag will be discussed next and it will be shown how difficult it is to estimate a time lag between measurements made at different stations. In the end, no calculated time lag was used in the comparison of the number fluxes. Time series of the NF and rainrate have been evaluated visually and a time lag has been introduced in cases where this seemed to be necessary to achieve a good correspondance of the data at the two stations.

For a time lag due to the fall time between two stations, two cases may occur. Either, measurements are compared between the top and the middle station, or measurements are compared between the middle and the bottom station. In both cases, however, the vertical distance is nearly equal and is somewhat less than 600 m. Now, an estimation has to be made about the fall time of hydrometeors falling 600 m. First, it has to be considered that parts of the melting layer may be between the two stations in question, i.e. fall velocity of hydrometeors may be quite low. Second, and much more important is the fact that the contribution of small particles to the NF exceeds the contribution of large particles. Much more small particles are present in precipitation than large particles. Thus, the fall velocity of the small particles determines the time lag of observations at two vertically displaced stations. The fall velocity of small particles may be assumed to be less than 1 to 2 m/s, depending whether ice particles or raindrops are considered. Therefore, a time lag of 5 to 10 minutes due to the fall time of hydrometeors has to be introduced when comparing number fluxes.

A second source for a time lag may result from horizontal advection of precipitation structures. Although all three cases considered here are stratiform and showers are not observed, some variation of precipitation is always present. These structures may be assumed to travel with the environmentlal wind. Here it has to be noticed that such a variation of precipitation reaches a point \( P_1 \) at the ground later than a point \( P_2 \) perpendicular above \( P_1 \) (see Fig. 4.8). The time lag \( \Delta t \) by which observation of this variation at \( P_1 \) and \( P_2 \) are shifted depends on the velocity of the wind \( (\bar{v}_w) \), the fall velocity of hydrometeors \( (\bar{v}_f) \) and the vertical distance between \( P_1 \) and \( P_2 \) (\( \Delta h \)).

The three stations along Mt. Rigi are set up more or less from West to East (Fig. 4.1). Thus, if winds are blowing from West the effect of advection reduces the effect of the fall time which was calculated to be at least five minutes. To calculate the effect of advection,
4.4. Overview of the three cases

the wind velocity has to be known. Wind velocities have been measured at all three stations along Mt. Rigi, but all these measurements were done close to the ground and do certainly not represent the winds within the free atmosphere. No soundings have been made. The only source of wind-data within the free atmosphere are VAD-products (Velocity Azimuth Display, Browning and Wexler, 1968) of the C-band Doppler radar in Zürich, 40 km away. These are not much of use because of the large distance.

However, on 19 March, winds were blowing very steady in Zürich during several hours with windspeeds of 5 m/s and a direction of WSW. Assuming the same wind was blowing in the region of Mt. Rigi, a time of 10 minutes is needed to travel the distance of 3 km between the bottom and the middle station. This time equals more or less the time needed for hydrometeors to fall the vertical distance of 600 m. Therefore, it seems that no net time lag is needed to compare number fluxes at the bottom and the middle station for 19 March.

4.4.2 Time lag introduced to the three cases

During two of the three cases, wind was generally blowing from the West. The third case was dominated by winds from the North. Wind directions and wind speeds measured at the three stations along Mt. Rigi and with the C-band radar (VAD) are summarized in Tab. 4.2 and Fig. 4.9. It has to be noted, that the wind directions on Rigi Staffel were strongly influenced by the surrounding orography and buildings and thus not representative. Typical wind direction on Rigi Staffel throughout the winter was around 150° which corresponds to a wind uphill through a valley (see the map in Fig. 4.1). Winds from 210 to 270° were considerably shielded by two buildings about 10 m away.

For the two cases with mainly westerly winds a small time lag would be expected since the stations along Mt. Rigi are put up also from West to East. Actually, a time lag of one minute
Table 4.2: Wind directions and wind speeds measured at the three stations along Mt. Rigi and with the C-band radar in Zürich. Wind direction 0 is North, 90 is East. No wind speeds are available for Greppen.

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<tr>
<td></td>
<td>Direction deg</td>
<td>Velocity m/s</td>
<td>Direction deg</td>
</tr>
<tr>
<td>VAD</td>
<td>235</td>
<td>12.5</td>
<td>360</td>
</tr>
<tr>
<td>Greppen</td>
<td>180</td>
<td>-</td>
<td>10</td>
</tr>
<tr>
<td>Seeboden</td>
<td>260</td>
<td>7</td>
<td>350</td>
</tr>
<tr>
<td>Staffel</td>
<td>170</td>
<td>4.5</td>
<td>300</td>
</tr>
<tr>
<td>General wind-direction</td>
<td>South-West</td>
<td>North</td>
<td>West</td>
</tr>
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</table>

Fig. 4.9: Wind directions measured at the three stations along Mt. Rigi and with the C-band radar in Zürich.

was found for the case of 14 February. The case of 19 March even seems to work without a time lag. Both cases show quite similar features in the time series of NF and rainrate. Thus, matching the time series and determining a possible time lag is not critical.

The case of 15 February is much more complex. Although precipitation is extremely uniform as can be verified with the RHIs, the time series of NF and rainrate of the bottom and the middle station show less correspondance than for the two other cases. A time lag of one minute was finally determined. This seems to be quite short since the wind is blowing from North, perpendicular to the connection line between the bottom and the middle station. All attempts to introduce a larger time lag resulted in badly mismatched time series of rainrate and were thus abandoned.
Chapter 5

Discussion

5.1 The largest particle measured with the optical spectrometer

The three additional cases show that the largest particles throughout the melting layer are observed approx. at the maximum of radar reflectivity (Fig. 5.1). For two cases maximum particle sizes are observed somewhat below the maximum of radar reflectivity, for one case somewhat above the maximum of radar reflectivity. Typical vertical width of the bright band is 200 to 600 m. Thus, maximum particle sizes are observed within 100 m of the maximum of radar reflectivity.

Some authors give maximum hydrometeor sizes within the melting layer together with radar reflectivity profiles. Stewart et al. (1984) and Willis and Heymsfield (1989) report the largest particles about 200 m above the maximum of the bright band whereas Zrnić et al. (1993) report the largest particles a few hundred meters below the maximum of the bright band. All these authors do not have high vertical resolution of neither radar reflectivity nor particle properties. This explains the large scatter of the observations of the vertical position of the largest particle.

The present results and the results found in the above mentioned literature indicate that the largest particles are not necessarily coincident with the maximum radar reflectivity although sometimes this is believed (Hagen et al., 1993). The particle size is one important factor in the formation of the radar bright band but other effects may be of, at least, the same importance.

5.1.1 Maximum particle sizes on 15 February 1997

Maximum particle sizes on 15 February 1997 are almost constant above and within the upper part of the melting layer. Particle maximum sizes do not show a peak around $H = 0$, thus it
seems that no aggregation has taken place within $2 \leq H < 0$. This is not the case and will be discussed in Sect. 5.3. For the time being, Fig. 5.2 shows a picture taken at 10:41 which corresponds to the position of $H = 0.5$. This is the first picture in a row of 10 pictures taken during the whole event on 15 February 1997. Ice particles and snowflakes were caught on a piece of knitted black mohair wool and then photographed. This picture shows clearly that many aggregates are present but they are rather small (8 to 10 mm). Later, aggregates have become less frequent and large single crystals can be found. The large picture in Fig. 5.3 shows a photo taken at 12:18. This time is beyond the selected time interval 10:00-11:20 but the picture shows essentially the same as two pictures taken at 11:10 and 11:20 which are slightly out of focus. Many large single crystals, mostly needles and dendrites can be found. Some of them have sizes of several millimeters. Furthermore, some small aggregates consisting of a few crystals can be found. A few plates can be found as well as columns capped with dendrites. Fig. 5.3 shows some single crystals. It can be seen that the dendrites have no delicate structures. On the large picture many columns capped with dendrites are visible, sometimes with different sizes of the dendrites. These crystals have a column between the two dendrites as found by Iwai (1983). Similar crystals are reported by Jiusto and Weickmann (1973) where two dendrites are connected by a frozen drop. Although similar in shape, these two type of crystals have a very different development: while the columns capped with dendrites have grown by vapor deposition on an ice nucleus, the other dendrites
5.1. The largest particle measured with the optical spectrometer

![Approx. 2 cm](image)

Fig. 5.2: Picture taken on 15 February 1997 at 10:41. Many small snowflakes are visible. have grown on a rather large frozen cloud droplet serving as an "ice nucleus".

5.1.2 Temperatures within the melting layer

To give a relation of particle size with respect to the radar reflectivity profile is one possibility to study particle sizes within the melting layer, another would be to compare particle sizes with the temperatures within the melting layer.

Temperatures have been measured at all stations and can be shown in dependance of the position parameter $H$ for cases where the melting layer passed one station. In addition to the four cases introduced in Sect. 4.2, two more cases can be found. On 6 March 1997, the melting layer was at the middle station although it was predicted to be at the top station. Thus, the optical spectrometer was not operated at the middle station but on top of Mt. Rigi, even above the top station. However, meteorological parameters were recorded on Seebodenalp and reflectivity profiles were measured at the bottom station. Therefore, the temperature in dependance of the position parameter could be determined. The second case is 28 March 1997 with the melting layer passing extremely quickly (within 10 minutes) the top station. Hydrometeor data of the top and middle station do not correspond to each
Fig. 5.3: Picture taken on 15 February 1997 at 12:18. Single crystals and small aggregates are visible. Five sections of the picture with exceptionally nice crystals are enlarged. The drawing illustrates the structures of the columns capped with dendrites and the picture bottom right shows a photo taken through a microscope of a FORMVAR slide. The "plates" capping the columns are clearly recognizable as dendrites.

other and can thus not be used for the thorough evaluation. The temperatures can be used, however.

Fig. 5.4 shows temperatures in dependance of the position parameter $H$. Each dot represents a one-minute value:
5.1. The largest particle measured with the optical spectrometer

- At the top station of Mt. Rigi, the temperature is measured every minute with an ultrasonic anemometer.

- In the case of Mt. Ūetli, temperatures on the top station have been constructed of data measured at five automatic ANETZ-stations (of the Swiss Meteorological Institute). Four of them were within 12 km and one at 30 km distance to Mt. Ūetli. Data of these five stations are average values of 10 minutes. These data have been interpolated for the height of the top station on Mt. Ūetli and for every minute. These interpolated temperatures correspond very well to the temperatures measured with one-minute resolution at the bottom station of Mt. Ūetli. The average difference between the temperature at the two stations is 3.6°C with a standard deviation of 0.2°C.

- Temperatures at the middle station on Mt. Rigi stem from an automatic NABEL-station (see Sect. 4.1) and are average values of 30 minutes. Again, one-minute values are interpolated.

It seems that temperatures measured at the top station on Mt. Rigi (14 February and 28 March) are not consistent with the temperatures measured at the middle station and at Mt. Ūetli. Furthermore, it seems that the top of the melting layer corresponds well with the 0°C isotherm. This is a confirmation that the procedure which is used to identify the top of the bright band really picks the physical relevant height of the onset of melting. The maximum of radar reflectivity is observed close to 1°C and the bottom of the bright band is located between 1.5°C and 3°C. Thus, particles are completely melted at approx. 3°C. Nevertheless, this is only valid for stratiform precipitation with a bright band and at most moderate riming. Heavily rimed ice particles or graupel in convective precipitation can survive unmelted a fall into temperature regimes above 10°C or in summer thunderstorms above 20°C.

The generally good correspondance of the temperatures within the melting layer in dependence of the position parameter \( H \) is an evidence that the conception of the position parameter is physically reasonable.

Fig. 5.1 can be compared to Fig. 5.4 and it can be seen that the largest particles are observed between 1°C and 2°C. Magono (1953) and Hobbs et al. (1974) both observe the largest snowflakes at -1°C. Hobbs et al. use the Formvar technique to replicate snowflakes. This technique does not work properly at temperatures around 0°C. This might explain why aggregates are found to be quite small at temperatures above -1°C. Rogers (1974) measures snowflake sizes by catching snowflakes on a velvet cloth. With this method it is possible to sample snowflakes at any temperatures. He observes the largest snowflakes (40 mm) at 0°C though snowflakes with diameter of 20 mm are observed between 1°C and 2°C. At last, giant snowflakes (60 mm) are reported by Auer (1971) at a temperature of 0.5°C.
Fig. 5.4: Temperatures within the melting layer in dependence of the position parameter $H$

5.1.3 Differential reflectivity observations within the melting layer

Another interesting method to possibly determine the position of the largest particles is to measure the differential reflectivity $Z_{DR}$ with a linear polarized radar. $Z_{DR}$ depends on the axial ratio of hydrometeors (Rinehart, 1991). Its value is zero if the cross-sectional area of the hydrometeor seen by the radar is circular and larger than zero if the area seen by the radar is oblate with the large axis oriented horizontally. This applies only to scanning radars viewing the hydrometeors under a certain (preferably small) elevation angle. Usually, particles with small axial ratios (height to width) are large particles. For raindrops, a well-defined relationship exists between equilibrium diameter and axial ratio (Beard and Chuang, 1987) and thus $Z_{DR}$ can be used to identify large drops.

Maximum values of $Z_{DR}$ in stratiform precipitation are generally observed somewhat below the maximum of radar reflectivity (e.g. Steiner et al., 1991, Zrnić et al., 1993, Hagen et al., 1993, Russchenberg and Ligthart, 1996). Axial ratios of snowflakes are not explored up to now but high values of $Z_{DR}$ within the melting layer are usually attributed to oblate snowflakes (Russchenberg and Ligthart). In Sect. 2.2.1, a first attempt was made to evaluate axial ratios of snowflakes and Fig. 2.10 shows, indirectly, that the larger snowflakes are, the smaller their axial ratio becomes. Thus, maximum values of $Z_{DR}$ might indicate the largest particles within the melting layer. However, the findings of Sect. 2.3.1 apply only to dry
snowflakes. It might well be that melting snowflakes of smaller size have an axial ratio even smaller than the large dry flakes. Thus, the maximum of $Z_{DR}$ does not necessarily detect the largest snowflakes. Steiner et al. (1991), Hagen et al. (1993) and Russchenberg and Ligthart (1996) observe the maximum of $Z_{DR}$ at approx. $H = -0.5$ which is somewhat below the position of the largest particles in Fig. 5.1. Zrnić et al. (1993) find the maximum of $Z_{DR}$ 700 m below the maximum of radar reflectivity which corresponds to a position even below the bright band.
5.2 The largest melted particle

The largest melted particle has been indentified for the three additional cases according to the method described in Sect. 3.6.2. The results are depicted in Fig. 5.5. The left picture in Fig. 5.5 shows the results with different symbols for the 4 different cases. The first detectable completely melted raindrops (~ 0.2 mm) are observed at 0.1 < \( H < 0.3 \). Above the maximum of radar reflectivity (\( H = 0 \)), the size of the largest melted raindrops show not much scatter and lie between 0.2 mm and 0.8 mm. As melting progresses within the lower part of the melting layer (\( H < 0 \)), the scatter becomes larger. For the same position within the bright band, the range of possible sizes for the largest melted particle become larger and larger. This does not seem to be a consequence of the different cases and thus of possibly different conditions because the data points show a large scatter for all cases. One possibility might be that this large scatter is a result of different rainrates. Although the construction of the position parameter \( H \) eliminates the effect of the rainrate, data have been evaluated considering different rainrates. The right picture in Fig. 5.5 shows that the rainrate is indeed not responsible for the scatter.

![Fig. 5.5: Left: The largest melted hydrometeors are shown with different symbols for the different cases. Right: The same as the left picture but with different symbols for different rainrates.](image)

Up to now, the position parameter \( H \) was used without the consideration that its conception is better suited to describe microphysical observations within the melting layer than the vertical distance to the top of the bright band. The largest melted drop is now a good occasion to demonstrate the potentials of the position parameter \( H \). The left picture in Fig. 5.6 shows again the largest melted drop, now in dependance not of the position parameter \( H \) but of the vertical distance to the top of the bright band. The main features remain the
5.2. The largest melted particle

The differences between Fig. 5.5 (left) and Fig. 5.6 are the most marked for the smallest sizes. While melted diameters of less than 0.5 mm occur within a small range of $H$, they can be observed within a range of nearly 200 m which makes up about 1/3 to 1/2 of the melting layer. This demonstrates that the position parameter $H$ is qualified as a dimensionless number for the discussion of the microphysics within the melting layer.

5.2.1 Comparison with model results

The largest melted hydrometeor or, which is equivalent, the melt distance of a snowflake of known weight has not been measured in field experiments until now. Laboratory studies, however, do exist as well as theoretical calculations. An early theoretical calculation of melt distances is presented by Wexler (1955). The author assumes spherical snowflakes and melting is calculated without considering evaporation or condensation. Fig. 5.7 shows the same data set as Fig. 5.6 together with some data of different authors. Two sets of melt distances calculated by Wexler are printed in Fig. 5.7 with solid curves labeled A and B. Both curves are calculated for a lapse rate of 6°C/km. Curve A applies to aggregates consisting of dendrites whereas curve B applies to aggregates consisting of columns and plates. The results do not fit well to the observed melt distances which may be a consequence of the use of too simple assumptions to calculate the process of melting.

Mitra et al. (1990) observed melting snowflakes in a wind tunnel and used these experimental data to validate their melting model. One experimental data set that is presented in their work consists of about 45 snowflakes with sizes of about 10 mm and a density of 0.02 g/cm$^3$, melting in air with humidity of 90%. These snowflakes would have a melted diameter of 2.7 mm. The average melt distance of these snowflakes is determined to 317 m. The blue
cross in Fig. 5.7 marks the average value of the wind tunnel experiment. It does not fit very well to the observations of the field experiments. One problem might be that Mitra et al. state in the figure caption of their Fig. 3, where the melt distance was taken from, that the density of the snowflakes is 0.02 g/cm³. This implies that this value applies for all 45 snowflakes. Somewhere else, the authors state that most of the experiments were carried out with snowflakes of about 10 mm in diameter ... with densities of 0.005 to 0.02 g/cm³. If this is true, melted diameters are somewhere between 1.5 and 2.7 mm. A snowflake of 1.5 mm melted diameter with melt distance of 317 m would fit very well to the data. Another problem might be the temperature gradient. Due to the technical circumstances, the temperature gradient in the wind tunnel experiments was not constant. At the onset of melting, the temperature gradient was 1.25°C/100 m, at the end 0.36°C/100 m. Although no soundings have been made during the experiments at Mt. Üetli and Mt. Rigi, an average temperature gradient can be deduced by considering Fig. 5.4. Here, the temperature difference between the top and the bottom of the bright band is approx. 2°C. The average width of the bright band is approx. 400 m. This would yield a temperature gradient of 0.5°C/100 m which is lower than the value of Mitra et al. Thus, the average snowflake of Mitra et al. melts too quickly and the melt distance is too short.

Fig. 5.7: Comparison between experimentally obtained melt distances with literature. The tiny black dots represent the same data set as in Fig. 5.6. The two curves labeled A and B are data taken from Wexler (1955). The blue symbols are data taken from Mitra et al. (1990) and the red, green and yellow dots are data taken from Matsuo and Sasyo (1981). The four lines are best fits whereby the black line represents the data marked with the black dots.

In addition to the wind tunnel data, Mitra et al. show calculated values for melt distances of snowflakes with diameters of either 5 mm or 10 mm and a constant density of 0.02 g/cm³. Furthermore, the temperature gradient and the humidity of the ambient air is varied. Results of these calculations are marked with numbered blue dots in Fig. 5.7. The symbols number 1 to 4 are all calculated with 100% humidity, number 5 with 95% humidity and number 6 with 90% humidity. Number 1 and 2 use a temperature gradient of 1°C/100 m, the rest of 0.6°C/100 m. These last four data points, calculated with a temperature gradient close to 0.5°C/100 m, fit very well to the field observations.
Another extensive data set of model calculations of melt distances can be found in Matsuo and Sasyo (1981). The authors calculate melt distances of snowflakes of various sizes and densities in air with three different values for humidity and a constant lapse rate of 0.6°C/100 m. These results are plotted in Fig. 5.7 with red, green and yellow dots for different values of humidity. The data with 100% humidity do not fit very well to the observations, whereas data with 90% and 80% humidity correspond rather well with the observations. Humidity throughout the observations at Mt. Üetli and Mt. Rigi was above 90%, mostly around 95%. Thus, it seems that the calculated melt distances of Matsuo and Sasyo are somewhat short.

5.2.2 Application of the largest melted particle

The largest melted hydrometeor within the melting layer is a very important quantity of models that are used to simulate stratiform precipitation. Various authors use models with microphysical parametrization for stratiform precipitation (Klaassen, 1988, Szyrmer et al., 1998, Göke and Waldvogel, 1997). All these models run with melting-routines and are used to calculate radar reflectivities. For both, melting and reflectivities, the largest melted hydrometeor is crucial. Usually, the melting-routine yields this largest melted hydrometeor which then is used to calculate radar reflectivities. Yet, no validation of this largest melted hydrometeor is done. This is mainly due to the rareness of appropriate observations.

A functional relation of the diameter $D_{\text{melt}}$ of the largest melted hydrometeor to the position parameter $H$ was constructed based upon the following considerations:

The function has to fulfill the two conditions:

\[
\lim_{H \to -1} D_{\text{melt}} = 0, \\
\lim_{H \to -\infty} D_{\text{melt}} = D_{\text{max}}
\]

where $D_{\text{max}}$ is the diameter of the largest raindrop observed. This requires a function of the general form of

\[
f(x) = \frac{1 - e^x}{1 + e^x}.
\]

Several constants can be introduced to control the precise shape of the curve. Adding 1 and scaling with $a$ shifts the values of $f$ from $[-1,1]$ to $[0,D_{\text{max}}]$. The parameter $b$ controls the steepness of the curve and $c$ inserts an offset to the x-axis ($e^{b(x+c)} = e^b e^{xc} = ce^{bx}$)

\[
f(x) = a \left( \frac{1 - c \exp(bx)}{1 + c \exp(bx)} + 1 \right).
\]

This curve is symmetric. The data require a nonsymmetric curve with a steeper part above $H \approx 0$ and a less steep part below $H \approx 0$. This asymmetry can be achieved by introducing another exp-function with two more constants:

\[
f(x) = a \left( \frac{1 - c \exp(bx) - d \exp(ex)}{1 + c \exp(bx) + d \exp(ex)} + 1 \right).
\]
This function can be transformed to the more simple form

\[ f(x) = \frac{a}{1 + c \exp(bx) + d \exp(ex)} \]

with the five fit parameters \( a, b, c, d, \) and \( e \). This form is too complicated and thus not adequate to fit data with much scatter. The introduction of \( d \exp(ex) \) results not only in an asymmetry but also in a shift of the x-axis, thus \( c \) and \( d \) are chosen to be 1 to achieve a more simple form. With this simple form fitting was performed together for all data from all four cases and the fit yields the following relation:

\[ D_{\text{melt}} = \frac{1.73}{1 + \exp(10.18H) + \exp(2.05H)} \]

where \( D_{\text{melt}} \) is the diameter (mm) of the largest melted hydrometeor at the position \( H \). The fitted curve is printed with a solid line in Fig. 5.5.
5.3 The number flux ratio

The right picture in Fig. 5.8 shows the number flux ratio summarized for all cases. The findings of Sect. 3.6.4 can also be observed for the three additional cases. The number flux ratio decreases rapidly within the upper part of the melting layer. The minimum of the NFR is reached at somewhat different positions $H$ for the four cases within the interval of $-0.5 < H < 0$. Fig. 5.8 left shows again the largest particles observed and it can be seen that the position of the maximum of the largest particles generally coincides with the minimum of the NFR. However, data of 15 February do not fit neatly to the other cases.

Fig. 5.8: Left: Same as Fig. 5.1. Right: The number flux ratio in dependance of the position parameter $H$ for all four cases. For the three cases 15 February, 17 February and 19 March the number flux density of the optical spectrometer is calculated by excluding the first two size classes as discussed in Sect. 3.6.4. For 14 February, both instruments, the optical spectrometer and the 2D-Video-Disdrometer have the same instrumental threshold. However, the NFR of all cases is not correct as soon as the instrument within the melting layer measures ice particles down to the smallest size classes. This might be the case for $H > 0.5$. 
The NFR on 15 February 1997

As for the evaluation of the largest hydrometeor, the case of 15 February 1997 stands somewhat apart from the other cases. The number flux ratio decreases rapidly within the upper part of the melting layer, shows a minimum and increases then again. On the whole, this course of the NFR corresponds to the observations made at Mt. Üetli. But when the NFR is inspected closer, it seems that the values of NFR are shifted towards smaller values of $H$. The data of NFR of 15 February 1997 would much better fit to the other cases if they were shifted by $\Delta H = 0.5$ "upwards".

To find a possible explanation, it has to be remembered that the position parameter $H$ gives the position of the optical spectrometer in relation to the radar bright band. Radar profiles are measured at the bottom station in a distance of 3 km to the middle station. The wind was blowing on 15 February 1997 at all stations except the top station on Mt. Rigi (which is not representative) from North (Fig. 4.9). If the direction of the wind is considered and the cold air is assumed to form a front perpendicular to the wind direction, the middle station is 1 km ahead of the bottom station. Therefore, cold air would reach the middle station earlier than the bottom station. From the transformation of time to $h$ it follows that a shift of $\Delta H = 0.5$ would correspond to a time difference of six minutes: If the NFR of a given time is related to a position parameter $H$ extracted from the radar profiles measured six minutes later, the NFR of 15 February would correspond perfectly to the other cases.

A possible time shift of five minutes may be estimated when the wind velocity of 3 m/s measured at the middle station and the distance of 1 km is considered. This is in a very good agreement to the required time shift of six minutes, but most probably a higher wind velocity should be used in this calculation since wind measurements at the middle station are made close to the ground. However, it seems that a certain time shift and hence a shift of $H$ might be appropriate for this case.

A possible mismatch of radar profiles and hydrometeor measurements have also to be considered for the other cases. For the case at Mt. Üetli, no such problem arises since the descent of the melting layer was very slow. On 14 February 1997 the melting layer did descend only marginally and hence there are no problems. At last, on 19 March 1997 the melting layer descended, but much slower than on 15 February 1997. Again, there are no problems.

Next, the validity of the three assumptions stated at the end of Sect. 3.7 will be tested with the results of the additional three cases.

### 5.3.1 Assumption 1: One snowflake yields one raindrop

There is only one more case that can be used to test this assumption. On 15 February, the NFR is 1.8 at $H = 1$. This is most probably not true. Just above, it was discussed
that the position parameter $H$ might be determined for this case too small. Furthermore, it has to be considered that small snowflakes taken into the sum of the number flux at the optical spectrometer might have melted diameters below the instrumental threshold of the disdrometer. The same correction as was used for the case of Mt. Üetli can be applied for this case, i.e. not only the two smallest size classes have to be excluded in the summation of the number fluxes but the fours smallest ones. With this correction, the NFR decreases to 1.5 at $H = 1$ which is still not consistent with the assumption. But at $H = 0.7$, the corrected NFR is equal to one. If there is really a time shift in the measurements, a shift of the position by $\Delta H = 0.3$ would be reasonable.

It seems that the assumption "one snowflake – one raindrop" may also hold for the case of 15 February 1997. The assumption is far from being proven, this would need additional cases, but it seems to be strengthened that on the average, one snowflake above the melting layer yields one raindrop just below the melting layer.

5.3.2 Assumption 2: Aggregation dominates over breakup within the upper part of the melting layer

Aggregation is the main process of growth of the size of hydrometeors in precipitation with low liquid water content above the melting layer (Houze, 1997, Barthazy et al., submitted). Aggregation is effective at temperatures as low as -20°C and generally the efficiency of aggregation increases with rising temperatures. However, aggregation seems to be rather effective at temperatures of the growth regime of dendrites around -15°C which is confirmed by observations of large aggregates at these temperatures (Magono, 1953, Rogers, 1974). As ice particles and snowflakes are close to the melting layer, aggregation increases again and particles tend to grow rapidly between about -6°C and 0°C. The enhanced efficiency of aggregation in this temperature regime is a consequence of a quasi-liquid layer forming on the surface of ice particles (for details see Pruppacher and Klett, 1997). This quasi-liquid layer forms as a result of the release of latent heat when crystals are growing by vapor deposition. Within the melting layer, aggregation is believed to be active as long as the snowflakes are not completely melted.

Although the position $H$ of the minimum of NFR varies slightly from case to case it still seems to be obvious that aggregation is effective above and within the upper part of the melting layer. This leads to the large snowflakes observed on 17 February 1995 and 14 February 1997 and the small number flux ratio within the melting layer observed for all four cases.
Aggregation on 15 February 1997

Again, the case of 15 February needs some further explanations. The number flux ratio decreases rapidly from $H = 2$ to $H = -0.5$ by a factor of 30 which could be interpreted as aggregation. In contradiction to this stands the nearly constant size of the largest snowflake. Here, no problems arise concerning the position parameter $H$ since a possible shift $\Delta H$ would apply to both, the NFR and the maximum sizes. A possible explanation for the rising NFR and the constant sizes can be found when crystal types are evaluated. On 15 February 1997 many needles were observed. Fig. 5.9 shows an aggregate with many needles caught on a FORMVAR coated slide. Needles were only occasionally observed in the other cases while on 15 February 1997 they may have had a portion of 30 to 50%. Literature about aggregates of needles is scarce, but it seems that although needles have a great tendency to form aggregates (Jiusto and Weickmann, 1973), they tend to stay small (Hobbs et al., 1974). This would explain the observed decrease of the NFR and the small maximum particle sizes. Fig. 5.10 shows the size distribution of snowflakes measured with the optical spectrometer at $H = 0.6$, where particle sizes are maximal and at $H = 2$ about 300 m above the melting layer. As expected if aggregation is effective, the size distribution at $H = 0.6$ shows much less small particles and more large snowflakes than the size distribution at $H = 2$.

One last point remains unexplained. The NFR of 15 February has its minimum at $H = -0.5$.

---

Fig. 5.9: **Left:** A rather large aggregate consisting of many needles and other crystals can be seen. **Right:** A small aggregate consisting of a few needles is shown. This small aggregate was found right below the large aggregate.
5.3. The number flux ratio

but at this position, maximum particle sizes are 3 mm which indicate that all particles are most probably completely melted. These two facts do not fit well to the assumption that aggregation dominates as long as the NFR decreases.

5.3.3 Assumption 3: Breakup dominates over aggregation within the lower part of the melting layer

The NFR increases within the lower part of the melting layer for all four cases. At the bottom of the melting layer the NFR is equal to 1 for two of the three cases with data available. This signifies that the total number of raindrops does not change when falling 450 m (Mt. Uetli) or 600 m (Mt. Rigi), respectively. This seems to be reasonable for stratiform precipitation with humidity near 100%. The case of 14 February lacks data around \( H = -1 \) and the case of 15 February does again not fit well to this concept if no shift of \( H \) is assumed.

The increase of the NFR is assumed to be due to the breakup of melting snowflakes. Mitra et al. (1990) have shown that the size of melting snowflakes remain largely unchanged until the melting process is advanced. They report an ice frame within the melting flake. Spontaneous breakup of melting snowflakes is not observed as a rule. However, they report breakup particularly when flakes with a strongly asymmetric mass distribution, held together by a few crystal branches, melt. Spontaneous breakup of such flakes may well take place in natural snow. Fig. 5.11 shows pictures of three snowflakes caught on a piece of fur. All of them show very delicate structures with only a few crystals connecting two parts of the snowflake.

Fig. 5.10: *Snow size distribution measured above the melting layer at \( H = 2.0 \) and within the upper part of the melting layer at \( H = 0.6 \). The spectra are averaged over five minutes.*
Fig. 5.11: Three dendritic snowflakes. The arrows point to delicate structures which could disrupt when melting and thus lead to breakup of the snowflake.

These snowflakes have not yet begun to melt but it seems to be likely that if the delicate connections marked with arrows in each flake melt, the snowflakes break up spontaneously.

Spontaneous breakup of natural snowflakes within a wind tunnel is observed by Fukuta et al. (1986). The authors describe the breakup process as rather dramatic: When snowflakes ... melted, they frequently disrupted. ... This occurred as if there were a small explosion in the snowflake. The temperature within the wind tunnel was 2 to 4°C and humidity was 90 to 100%. Melting of snowflakes (diameter not reported) took at most 20 seconds under the temperature and humidity ranges of this study. The observed "explosions" might be due to the high temperatures and the resulting quick melting of the flakes. Such melting behaviour is probably not occurring or at least not occurring frequently under natural melting conditions.

Breakup of ice crystals is observed by Oraltay and Hallett (1989) in air with relative humidity below 70%. Dendritic crystals are observed to produce an average number of 30 pieces per 5 mm crystal. Breakup of ice crystals is also used by Hobbs et al. (1976) to explain the observed large number of ice crystals within clouds which exceed the number of ice nuclei sometimes by five magnitudes.

Although several authors observe or believe in breakup of ice crystals, this is most likely not the process that leads to the observed increase of the number flux ratio within the lower part of the melting layer for the cases at Mt. Üetli and Mt. Rigi. The largest melted hydrometeor has at this height a diameter of at least 0.5 mm. Thus, most single crystals are already melted and cannot contribute to the rising number flux by breaking up. The number flux has to rise as a consequence of breakup of the larger snowflakes.

Spontaneous breakup is one possibility for the breakup of snowflakes. Within natural precipitation, collisional breakup may become important. The range of the particle fall velocities within the lower part of the melting layer is due to the partly melted particles larger than
5.3. The number flux ratio

in the upper part. It could be seen in the case of Mt. Úetli that hydrometeor fall velocities within the upper part of the melting layer (Fig. 3.12, 11 spectra of averaged fall velocity from 22:30 to 01:00) are not different from the typical fall velocity of snowflakes. This observation is consistent with the results of the wind tunnel experiments by Mitra et al. (1990). They observe that the fall velocity of a melting snowflake does not change by more than about 10% until 50% of the mass of the snowflake has melted. Within the lower part of the melting layer, the melted fraction of the snowflakes increases rapidly which is followed by a rapid increase of fall velocities. In the middle of the lower half of the melting layer \((H = -0.5,\) at 21:30 in the case of Mt. Úetli) the range of fall velocities has become rather large. The largest completely melted particles have fall velocities between 5 and 6 m/s while some of the snowflakes still have fall velocities between 1 and 2 m/s. This large range of fall velocities together with the rather large size (> 10 mm) of the melting snowflakes may well result in a high rate of collisions. The available energy in case of a collision is larger in the lower part of the melting layer than in the upper part and thus collisions could induce more breakup of snowflakes within the lower part of the melting layer than in the upper part. Furthermore, most of the large snowflakes are in an advanced state of melting and may breakup upon collision easier than dry snowflakes.

The development of the fall velocity of hydrometeors within the melting layer is for all three additional cases similar to the case at Mt. Úetli and are thus not presented separately.
5.4 Qualitative estimation of aggregation and breakup efficiencies

This section is devoted to a first attempt of a qualitative estimation of aggregation efficiencies. The case of 15 February 1997 is used since for this case measurements of snow are available above the melting layer up to a position of $H = 3$. This estimation is done solely with the data of the optical spectrometer and the radar. Therefore, no problems arise related to mismatches in time for two instruments or to different thresholds of instruments as was the case in calculating the NFR. However, a shift of $H$ might be possible as it was discussed in Sect. 5.3.

5.4.1 Size distribution of snowflakes

Instead of printing a serie of snow size distributions, $N_0$ and $\Lambda$ have been calculated for every minute, assuming an exponential size distribution. The results are shown in Fig. 5.12. No values of $N_0$ and $\Lambda$ were calculated before 10:35 (or below $H = 0$) for at this time the size spectra become significantly different from an exponential size distribution due to melted particles and calculating $N_0$ and $\Lambda$ does not make sense any more.

Fig. 5.12: The intercept and slope parameter of an exponential size distribution fitted to the measured snow size spectra on 15 February 1997.
5.4. Qualitative estimation of aggregation and breakup efficiencies

The evolution of the size spectra is monitored by the parameters $N_0$ and $\Lambda$. The intercept parameter $N_0$ is decreasing from about 15'000 at $H = 2.5$ to less than 1'000 at $H = 0$. The slope parameter $\Lambda$ decreases also, from about 1.5 at $H = 2.5$ to less than 1. The changes of $N_0$ and $\Lambda$ indicate that the size spectra become less steep as snow drifts downwards.

A similar evolution of $N_0$ and $\Lambda$ has been observed by Stewart et al. (1984) for a penetration of the melting layer of a stratiform precipitation event. They depict their measurements in relation to the ambient temperature and show the evolution of hydrometeor size spectra from -15°C to +10°C. On 15 February 1997, the position of $H = 2.5$ corresponds to -1.5°C and the position $H = 0$ to +1.2°C. The evolution of the size spectra shown by Stewart et al. is somewhat slower since they observe $N_0$ to decrease by a factor of about 3 between -1.5 and +1.2°C whereas on 15 February 1997 $N_0$ decreases by more than a magnitude. The decrease of $\Lambda$ is also stronger for the case at Mt. Rigi than for the case reported by Stewart et al.

Lo and Passarelli (1982) observe the evolution of snow size spectra with airborne instruments above the melting layer. They report that the slope parameter $\Lambda$ of the snow size distribution decreases as snow is drifting downward and evolves towards a certain limit which is not observed to be surpassed. The minimum slope parameter the authors observe is $\Lambda = 1$ mm$^{-1}$. This value is in good agreement to the snow spectra measured on 15 February 1997.

The behaviour of the snow size distribution as measured on 15 February 1997 implies that aggregation is effective above and within the upper part of the melting layer. This result was already achieved by comparing particle number fluxes at the optical spectrometer and at the disdrometer. But no prediction could be made whether aggregation is enhanced within the melting layer or not. An attempt to estimate the aggregation efficiency above $H = 0$ will be made in the next section.

5.4.2 A qualitative estimation of aggregation efficiencies

In Sect. 4.4 it was shown that the NF at the optical instrument increases rapidly after 10:30 while the NF at the disdrometer remains more or less constant. The left picture in Fig. 5.13 shows again the NF at the optical spectrometer in dependance of the position parameter $H$. But here, not one-minute values are shown but the average number flux for a certain interval of $H$. 22 equidistant intervals of $H$ are created with the width of $\Delta H = 0.2$ and the mean values of $H$ shown below:

<table>
<thead>
<tr>
<th>Interval No.</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>...</th>
<th>20</th>
<th>21</th>
<th>22</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean value of $H$</td>
<td>3.0</td>
<td>2.8</td>
<td>2.6</td>
<td>2.4</td>
<td>2.2</td>
<td>...</td>
<td>-0.8</td>
<td>-1.0</td>
<td>-1.2</td>
</tr>
</tbody>
</table>

As it can be seen in Fig. 5.13 (left), the decrease of NF from $H = 3$ to $H = -0.5$ is rather smooth. For each interval, the NF was averaged over all one-minute values with according
values of $H$. This meant averaging over 1 to 11 minutes whereby averaging was done mostly over 2 to 4 minutes. Furthermore, this time the number flux is calculated over all size classes of the optical spectrometer. Since no comparison is made with the disdrometer, there is no need to exclude the two smallest size classes of the optical spectrometer in summing up the number flux. Thus, the NF shown in Fig. 5.13 (left) is somewhat higher than the NF of the optical spectrometer shown in Fig. 4.5.

Fig. 5.13: Left: The number flux measured at the optical spectrometer on 15 February 1997. The height is divided up into 22 intervals with the same width $\Delta H = 0.2$. Average NF is calculated for every interval. Right: The flux ratio (see text) calculated for every interval is shown. The solid line is drawn by hand to depict the decrease of the FR between $H = 3$ and $H = -0.4$.

To proceed, the following assumption is made: There is no change of the precipitation from one interval to the next related to any meteorological parameter except a decrease of the particle number flux. The time interval this assumption applies for is for most of the two succeeding intervals less than 10 minutes. A flux ratio is then calculated for every two
adjoining intervals:

\[
\text{flux ratio}_n = \text{FR}_n = \frac{\text{average number flux in interval } n - 1}{\text{average number flux in interval } n}
\]

If the value of the FR is larger than one this signifies that the number flux of snowflakes falling down has increased from one interval to the next. If the value is smaller than one, the number flux has decreased. The right picture in Fig. 5.13 shows this flux ratio in dependence of the position parameter \( H \). At \( H = 2.8 \) the FR is close to one indicating that the NF does not change significantly from one interval to the next. But as snow is drifting down, the FR decreases slowly. At \( H = 1 \) about 30% less particles are counted than at \( H = 1.2 \). This is a significant decrease of number flux. The smallest value of the flux ratio is reached at \( H = -0.4 \). Here, almost half of the particles get lost between \( H = -0.2 \) and \( H = -0.4 \). Below \( H = 0.4 \), the flux ratio changes to values larger than 1. At \( H = -0.8 \) suddenly two times more particles are present than the interval before at \( H = -0.6 \). This sudden increase of the NF comes to a stop and at \( H = -1.2 \) the same number flux can be observed as at \( H = -1.0 \).

The interpretation of the right picture in Fig. 5.13 is that aggregation is active above and within the upper part of the melting layer. The efficiency of aggregation increases as snowflakes are falling down and shows a maximum at \( H = -0.4 \). At this height, 40% of the particles get lost within a time interval of four minutes and within a vertical distance of approx. 80 m. Then suddenly this tendency is reverted and at \( H = -0.8 \) two times more particles are observed than one interval higher up within a time interval of 6 minutes and again a vertical distance of approx. 80 m. This sudden increase of particle number flux clearly indicates efficient breakup.

At last, it has to be mentioned that the relation of the flux ratio in Fig. 5.13 to the position parameter \( H \) might be mismatched. As stated in Sect. 5.3, \( H \) might have to be corrected towards higher values. If this applies, the minimum of the FR and hence maximum aggregation efficiency would not be found around \( H = -0.4 \) but somewhat higher up and closer to the height of the maximum reflectivity.

5.4.3 The case of Mt. Üetli

The same methods as have been shown above for the case of 15 February 1997 have been used to estimate aggregation and breakup efficiencies for the case at Mt. Üetli. Fig. 5.14 shows the number flux and the flux ratio calculated for equidistant intervals of \( H \). The only difference in the evaluation between this case and the case of 15 February 1997 is that the intervals are chosen with a width of \( \Delta H = 0.1 \) since much more data are available. In the case of Mt. Üetli, no significant aggregation can be found between \( H = 1 \) and \( H = 0.5 \). The minimum of the FR, however, has about the same value (FR=0.6) as in the case of 15
February 1997. Again, about 40% of the number flux is lost, this time within 16 minutes and within a vertical distance of approx. 60 m. The minimum of the FR is observed for this case at the maximum of radar reflectivity. Below $H = 0$, again breakup can be observed producing two times more particles than have been observed one interval higher up.

![Graph](image)

Fig. 5.14: **Left:** The number flux measured at the optical spectrometer on 17 February 1995. The height is divided up into 26 intervals with the same width $\Delta H = 0.1$. Average NF is calculated for every interval. **Right:** The flux ratio (see text) calculated for every interval is shown. The data missing are due to a short circuit.

The reason why no significant aggregation can be observed within the upper part of the melting layer in the case of Mt. Üetli might be that the precipitation rate is not constant. In contrast to the case of 15 February 1997 where the precipitation rate was rather constant, in the case at Mt. Üetli the rainrate did increase during the begin of the experiment and decrease during the end of the experiment. Around both positions, $H = -1$ and $H = 1$, the rainrate was rather low. If the rainrate at these two positions had the same value as around $H = 0$, the NF would be higher at these positions. Then, probably, a marked decrease of the number flux could have been observed between $H = 1$ and $H = 0$ which would indicate
aggregation and at the bottom of the bright band \((H = -1)\) the NF would remain at a constant value indicationg the end of the breakup process.
Chapter 6

Conclusions

The present work is based upon several experimental observations made in stratiform precipitation in winter time. Profiles of microphysical properties of stratiform precipitation and especially of the melting layer were obtained by placing instruments at several stations along the steep slope of two different mountains. In cases with a descending melting layer passing one station, measurements within the melting layer could be made. The strict criteria in selecting good cases did leave no more than four cases to be included in this thesis. The thorough evaluation of these four cases has shown that ground-based measurements of microphysical properties within the melting layer of stratiform precipitation can yield information that cannot be obtained by airborne measurements. The evaluation led to the following findings and assumptions:

1. A dimensionless position parameter $H$ was constructed for stratiform precipitation using the top, the maximum and the bottom of the bright band obtained from vertical radar profiles. This parameter was shown to be suited better for the discussion of microphysical properties within the melting layer than the vertical distance to the 0°C isotherm or to the top or the melting layer, which is equivalent.

2. A classification scheme was developed based upon RHI products of a scanning radar which can be used to separate uniform precipitation in space and time from any other precipitation type.

3. The largest snowflakes were observed within 100 m of the height of the maximum of radar reflectivity. This height was found to correspond to temperatures between 0.5 and 1.5°C. At the bottom of the bright band, the temperature was approx. 2.5°C. Here, all ice particles and snowflakes are completely melted in stratiform precipitation.

4. The largest melted hydrometeor in dependance of the distance to the top of the melting layer or, which is equivalent, the melt distance of a snowflake of given mass can be measured with ground-based instruments. The measurements were found to be in good
agreement with theoretical calculations of melt distances. Furthermore, it was shown that the position parameter $H$ is better suited to depict the largest melted hydrometeor than the vertical distance to the top of the melting layer and a functional relationship was deduced.

5. By comparing number fluxes (number of hydrometeors falling through one square meter within one minute) at the top and at the bottom of the melting layer, the assumption was made that, on the average, one snowflake yields one raindrop.

6. By comparing number fluxes in the upper part of the melting layer and in rain below the melting layer, aggregation seems to be the cause of a rapid decrease of the number of particles within the upper part of the melting layer. This strong aggregation leads to the large snowflakes observed in the middle of the melting layer.

7. By comparing the number flux in the lower part of the melting layer and in rain below the melting layer, breakup seems to be responsible of a rapid increase of the number of particles within the lower part of the melting layer. Spontaneous and collisional induced breakup seems to occur frequently within the lower part of the melting layer where the range of fallspeeds of the hydrometeors is rather large.

8. The efficiency of aggregation seems to be more and more enhanced as snowflakes are drifting downwards and entering the melting layer. The highest efficiency of aggregation can be found within the melting layer close to the maximum of radar reflectivity.

The assumptions made above, especially those regarding aggregation, breakup and the efficiency of aggregation do not stand on a firm base. Too few and too different are the four cases to serve as a good foundation of experimental data. More observations are required and should be done, since ground-based experiments are, as was shown in this work, the only possibility to measure quantities such as the fall velocity or the number flux of hydrometeors.
Chapter 7

Outlook

The results of the experiments discussed in this thesis show just one part of the research done. Much more can, will and should be done with the enormous amount of data collected in these extensive field campaigns. Furthermore, improvements are desirable, both of the instrumentation and of the design of the experiments. This chapter will show some possible developments.

7.1 A new instrument

To measure automatically good quality size distributions of snow, particle measuring instruments working on optical principles are needed. Since exact relationships of particle diameter to particle fall velocity do not exist for snowflakes, these optical instruments not only have to measure the size of a snowflake but also its fall velocity. The deficiency of the optical spectrometer used throughout the experiments was its inability to measure exact fall velocities. The principle of the 2D-Video-Disdrometer is perfectly suited to measure fall velocities of snowflakes but in practice, the instrument could not meet the high expectations. The main disadvantage was the direction of the two light beams, perpendicular to each other. Snowflake images were too different from these two sides to be matched reliably which is the prerequisite for velocity measurements.

The problems with measuring exact fall velocities was one reason why the main findings of this thesis are based upon number flux densities which are independent of the measured particle fall velocity.

The consequence of the experiences made in the winter season 96/97 was the initiation of the development of a new optical instrument. Its main principle is the same as the one of its predecessor, the optical spectrometer used since many years, with the only difference of using two light beams parallel to each other vertically offset by approx. 10 mm. First experiments have been conducted with this new instruments since January 98. Although
the evaluation software has not yet been developed, the results are promising.

### 7.2 New experimental setup

The experimental setup at Mt. Rigi had the main disadvantage that the stations were fixed. If the melting layer did not pass either the top or the middle station, no measurements could be made. Stratiform precipitation lasting for at least 2 hours with the melting layer descending proved to be quite rare. Some events were probably missed due to inexact forecasts, but out of the three perfect cases in two the melting layer did not pass an equipped station. Thus, the thought arose that more data could be collected if the stations were not fixed but mobile. Two scenarios were examined, one with instruments mounted on a car driving up a steep road and one with instruments mounted on a cable car. Possible locations were found for both methods. The method with the car suffered mainly from finding a road, ridable in winter, leading up to a convenient height within a short distance. The best road that could be found climbed from 600 m ASL to 1400 m ASL within 3.5 km. 1400 m ASL did not seem to be high enough to be sure to have snow. The method with the cable car seemed to be more promising, moreover since a location could be found with a vertical climb of 1100 m to 1900 m ASL within a horizontal distance of 1.5 km.

The decision was reached to use the cable car and the winter campaign 97/98 was planned at Lithal, involving the cable car of the power station Kraftwerke AG Linth-Limmern. This cable car is only used by the workers of the power station and was thus almost always available for experiments. Again, the winter season proved to be very dry and the peculiar micrometeorology of the Linthal valley with unpredictable Föhn-effects added to this dryness. Furthermore, the weather forecasting seemed to be very difficult for this remote valley. At Linthal, the bottom station was equipped analogous to the other experiments. Meteorological instruments and the optical spectrometer were mounted on the roof of the cable car, Formvar and photographic facilities were set up within the cable car. No third station was used.

The data of this winter still have to be evaluated thoroughly to judge the usefullness of the setup with the cable car.

### 7.3 Particle shapes and fall velocities

Although fall velocity measurements of snowflakes have been done very painfully by many authors, the statistics are not very good with the number of snowflakes sampled being some tens to a few hundreds. Only automatic measurements with instruments capable to measure exact fall velocities for thousands of snowflakes will enhance the statistical significance of particle size to velocity relationships. Furthermore, fall velocity measurements of snowflakes
larger than 15 mm are rather scarce which is a significant gap.

The shape of snowflakes, especially their axial ratio (height to width) has not been investigated up to now, despite the fact that this could be interesting. Measuring fall velocities of snowflakes with optical instruments working with only one light beam is closely related to the axial ratio of the snowflake. Moreover, radar measurements with linear polarized radars are also related to axial ratios of hydrometeors which are up to now only known for raindrops.

7.4 Melting layer models

All data obtained in experiments are extremely valuable for modeling. Having a good set of experimental data is crucial in validating models. One promising possibility to use the high resolution sequence of snowflake number fluxes measured with the optical instrument is to calculate aggregation efficiencies. Breakup efficiencies are also a challenge, but since no measurements were made about the number of debris a snowflake is breaking up into, validation will be less reliable.
Appendix A

The classification of non-uniform precipitation

In section 3.3 a classification scheme has been established with the help of which stratiform and uniform precipitation can be distinguished from any other type of precipitation. As a contrast to the case shown there, two precipitation events will be presented in the same manner in this section, one stratiform with embedded shower cells and one thunderstorm.

A.1 Stratiform precipitation with embedded shower cells

On December 14 1996 precipitation was mainly stratiform but the appearance of the rainfall was by no means uniform. Shower cells with typical horizontal extensions of 10 km were passing by with only weak precipitation in between. The showers caused quite high rainfall rates of about 3 to 4 mm/h while between the shower cells rainfall rates were lower by about a magnitude. However, bright band structures were well developed within the most intensive shower cells as well as between them. Fig. A.1 shows an RHI. The shower cells and the bright band along the first 40 km are easily visible. Echo top is varying with the intensity of precipitation.

A time interval of 3 h 50 min with 24 RHIs have been chosen. Standard deviations and percentages of occurrences of standard deviations have been calculated according to Sect. 3.3 for 3-point-smoothing. The results are depicted in Fig. A.2. The most frequent standard deviations are below 2 dB, no values over 10 dB are observed except behind the ground clutter. However, picture 6 (Clutter or no precipitation) reveals the "patchy" structure of the precipitation. Behind Mt. Rigi (clutter at 40 km) precipitation was decreasing quickly and ahead of Mt. Rigi the RHIs have many pixels without precipitation. This time the wind was not blowing perpendicular to the radar beam and these are real pixels without
The classification of non-uniform precipitation

December 14 1996, 02:52

Fig. A.1: RHI of a stratiform precipitation with embedded shower cells.

precipitation. Nevertheless, Fig. A.2, with smoothing over 3 points, is not much different from the case presented in Sect. 3.3.

When performing a smoothing over 21 points the differences are apparent. Fig. A.3 shows these results. There are nearly no RHIs with pixels with standard deviations below 2 dB. Most pixels have standard deviations between 2 and 10 dB or fall into the category of no precipitation. The "darkest appearance" shows, at least ahead of Mt. Rigi, picture 4 (4--10 dB). This stands in contrast to the findings of the case in Sect. 3.3 and therefore a tool is created to discriminate, simply by looking at Figs. 3.8 and A.3, between uniform and showery precipitation with otherwise stratiform structures.

The investigated period shown here consists of almost four hours of precipitation. It has been shown that precipitation was showery. However, it still could be that precipitation during some shorter period within these four hours was uniform. The results above do not exclude this case. Maybe, if the time period had been chosen different, e.g. shorter, the result would have been different. But since this was not done, results are valid together for the chosen range and time interval.
A.1. Stratiform precipitation with embedded shower cells

Fig. A.2: Categories of standard deviations if smoothing is performed over three points.
The classification of non-uniform precipitation

Fig. A.3: Categories of standard deviations if smoothing is performed over 21 points.

start: date 961214 time 00:02
end: date 961214 time 03:52
total number of RHIs: 24
means calc. over # of points: 21
On 14 July 1995, a thunderstorm passed within 10 km of the C-band radar. A sequence of 13 RHIs have been taken within 100 minutes. One of them is shown in Fig. A.4. Reflectivities of more than 40 dBZ are reaching up to 10 km above ground level. Fig. A.5 shows the results if a smoothing over 3 points is applied to these RHIs. Surprisingly the most frequent standard deviation is less than 2 dB, even within the core of the thunderstorm. This shows that smoothing over 3 points does not yield any useful information about the type of precipitation. But strikingly different is Fig. A.6 where 21-point-smoothing has been applied. Virtually no standard deviations below 4 dB are observed. In contrast to the two other cases (uniform and showery stratiform precipitation) where no standard deviations above 10 dB have been observed, about 50% of the RHIs have standard deviations of more than 10 dB within the core of the thunderstorm. It is proposed that thunderstorms can be identified if picture 5 (> 10 dB) shows for 21-point-smoothing a percentage of occurrence of at least 25%.

Fig. A.4: RHI of a thunderstorm with heavy rainfall.
Fig. A.5: Categories of standard deviations if smoothing is performed over three points.
Fig. A.6: Categories of standard deviations if smoothing is performed over 21 points.
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Curriculum Vitae

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List of publication

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