Textural and mechanical variability of mountain snowpacks

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Chapter 1

Introduction

The variability of physical properties in porous media is an area of strong interest to the environmental science community. Most porous media occurring in nature show a distinct spatial variability. To have a tool in the mathematical treatment of this phenomenon the science of spatial statistics has been developed (Matheron, 1965). Cressie (1993) describes the methods and applications in his monograph. Fenton and Vanmarcke (1991) developed an efficient method for the simulation of random fields and their use in finite-element models.
Snow is in this respect a model material, typical to the properties of many different sedimented porous media. Snow is also a very special medium as it consists only of a single mineral, ice. This property may simplify many types of investigations but is at the same time sometimes cumbersome due to the inclination to rapid metamorphic changes.

The spatial variability of the snow is of prime interest to:

- Snow hydrology (generation of preferential flow path and water flow), (Carroll and Cressie, 1997)
- Snow chemistry (solute transport of adsorbed chemicals)
- Snow mechanics (development of simulation models used in avalanche forecasting and vehicle trafficability), (Smith and Sommerfeld, 1985; Conway and Abrahamson, 1988; Föhn, 1988; Jamieson and Johnston, 1993; Birkeland et al., 1995; Sturm et al., 1997)
- Snow ecology (food availability and resistance to movement), (Huggard, 1993; McRoberts et al., 1995)

In the following, the state of the art of methods available to investigate spatial variability of snow, and the controversies created by uncertain data (Smith and Sommerfeld, 1985; Conway and Abrahamson, 1988; Föhn, 1988; Jamieson and Johnston, 1993) are explained. The initial cause for spatial variability of snowpacks is created by the terrain and weather during snow precipitation and deposition. Subsequent precipitation events are forming layers. Layers deposited during stronger winds may change in their physical properties within a few 0.1 m. The structure and texture[1] of the snow crystals are then modified due to rapid thermodynamic processes, summarised under the term „metamorphism“. It is supposed that certain metamorphic processes, especially the formation of subsurface hoar, are highly variable in spatial extent (again a typical scale of a few 0.1 m is supposed, but this number is very uncertain). This causes many controversies in the avalanche forecast community, because the spatial distribution of so-called weak spots is extremely relevant to natural avalanche release. However, the investigation of these properties and spatial variation could not be answered with the detail necessary to apply the concepts of random fields. This was caused by a serious lack of an instrument resolving layers with an accuracy of about 1 mm that is at the same

[1] Texture is defined as the size, shape and arrangement of grains in a specimen of snow, structure defines the properties of the grains.

time field portable, rapid, and permitting a physical interpretation of the measurements. Johnson and Schneebeli (1998) and Schneebeli and Johnson (1998) developed the SnowMicroPen (SMP), a new type of a high resolution snow penetrometer. The displacement resolution of this instrument is 0.004 mm. The SnowMicroPen force signal can be interpreted and correlated to the texture of snow. This correlation is based on measurements of more than 40 well-defined laboratory snow samples. The correlation to snow texture was done by statistical methods (Pielmeier, 1998, Schneebeli et al. 1999) and by developing a micro-mechanical model (Johnson and Schneebeli, 1999). Parallel measurements in sieved and homogeneous natural snowpacks show excellent correspondence. This is an improvement in vertical resolution of about a factor 100. Two persons are able to measure more than 100 profiles (of 1.5 m depth) within a day, about 20 times faster than classical snow profiles. This instrument has therefore the potential to overcome the limitations typical to all formerly used methods and based on it a systematic spatial investigation of stratified snowpacks at different scales is now possible.

Table 1 presents an overview of available methods in snowpack investigations and their limitations. The methods are of varying relevance. The classical snow profile, based on the International classification for seasonal snow on the ground (Colbeck et al., 1990), has a high relevance today. It delivers an integral description although its objectivity and resolution are limited.

Instrument	Type of measurement	Objectivity	z-resol. [m]	horizontal-resol. radius [m]	Time [min]	Comment
Shear Frame	mechanical	High	0.01	0.1	20	destructive, only single layer
Stuff Block	mechanical	Medium	-	0.3	10	destructive, only first weak layer
Rutsch Block	mechanical	Medium	-	1.2	30	destructive, only first weak layer
Rammsonde	mechanical	Medium	0.04	0.02	20	
SnowMicroPen	mechanical	High	0.001	0.003	5	
Snow Profile	descriptive	Low-Med.	0.01	0.1	40	destructive
Translucent Profile	optical	High	0.005	0.005	40	destructive
Near-Infrared Profile	optical	High	0.005	0.005	40	destructive
Radar	electro-magnetical	High	0.05 ?	0.05	60	Interpretation difficult
Serial Sections	optical	High	0.00001	0.00001	50	destructive
Density		High	0.04	0.2	3	destructive

Of all methods, only the SnowMicroPen, has enough spatial resolution and is at the same time quick enough to measure the spatial variation of a snowfield. Due to the lack of appropriate field investigation methods, the spatial structure of the snowpack has only been investigated in a very limited way (Conway and Abrahamson, 1988; Föhn, 1988). None of the investigations were able to generate sufficiently large datasets to calculate the one- or two-dimensional autocorrelation functions. This information is a necessity to assume the correct random distributions of properties, as it is necessary for finite-element simulations of water infiltration, snow creep or avalanche formation. Hence, the spatial variability of snow texture and snow mechanical properties is badly understood. The development of more refined snowpack simulations will be of little scientific and practical value without more highly detailed investigations of the textural and mechanical properties and their variations.

The main goals of the thesis are to measure and analyse the micro structural and micro mechanical properties of the complete natural snowpack using the SMP, and to establish a systematic method for the investigation, analysis and interpretation of the small scale spatial variability of mountain snowpack properties. To meet these aims, the following steps have been taken:

- Application and development of interpretation methods of the SMP force-distance signal to field profiles.
- Comparison of measured snow micro properties with simulated snow micro properties.
- Characterisation of the properties of natural versus artificial snowpacks on ski slopes.
- SMP force signal calibration and improvements of the sensor technology for field applications.
- Systematic measurements, analysis and interpretation of the spatial variability of the textural and mechanical snowpack properties.

The thesis consists of this introductory chapter, five main chapters containing two published papers and three papers submitted for publication in reviewed journals, and the final synopsis chapter:

Chapter 2 is a literature review on the developments in snow stratigraphy over the last 150 years. The paradigms in snow stratigraphy shifted from a slope-scale, geomorphologic-sedimentologic approach in the beginnings to a one-dimensional approach to the physical and mechanical properties in “homogenous” layers in the 1940’s. The concept of well-defined layers gets accepted and this paradigm still dominates today. Avalanche formation is related to the spatial variation of snow properties. New instrumental developments show that the perceived strict layering may be a too simple model and modern snow stratigraphy has to integrate different scales. In this review the different directions taken in snow stratigraphy are followed and their suitability to meet the requirements in modern snow research is discussed.

Chapter 3 introduces the method of snow micro penetrometry and the texture index is applied for the first time to interpret a natural snow profile. Geostatistical analysis is carried out on SMP measurements of homogeneous snow samples. Spectral signal analysis is a suitable tool for the quality control of the SMP measurements.

The texture index is the fraction of mean grain size (mm) and density (kg m^{-3}) and therefore a direct index of the volume density of grain contacts. It is correlated to the coefficient of variation of the SMP force signal. The idea behind the texture index is that smaller spherical structural elements have a decreasing texture index, which is an indication of increasing textural stability.

Chapter 4 compares measured and modelled snowpack parameters. An empirically derived texture index is compared with a simulated texture index for laboratory experimental snow and natural snow profiles. The texture index was calculated throughout the development of three temperature gradient snow experiments from laboratory measurements of grain size and density. In a one-dimensional finite element snowpack model, the development of the mean grain size and density of these snow samples were simulated. It can be shown that the modelled texture index can predict the measured texture index. Three natural snow profiles from the Weissfluhjoch test field are compared with the texture index simulations. The latter comparison reveals shortcomings in the measurement of hardness and grain morphological parameters by classical methods. It also shows the dependency of a good physical model on quantitative process descriptions at the micro scale.

The classical snow profile is a one-dimensional record of the snow properties. A vertical profile wall is opened by digging a snow pit. The observer establishes the stratigraphy of the snowpack by sensing hardness differences by hand and inspecting textural differences by eye at a vertical line at the profile wall. For each layer the hardness and morphological snow properties are classified.

Chapter 5 is a study of the thermal impacts of ski-slope preparation in a sub-alpine ski resort in central Switzerland where artificial snow is produced. Soil temperature measurements and a characterisation of the snowpack properties were carried out. A numerical soil-snow-atmosphere transfer model with a new-implemented option to simulate the snowpack development on a groomed ski-slope was run for one of the ski-slope and off-piste sites. For the soil temperatures, the considerable differences in the snow thermal properties between ski-slope and off-piste showed to be less important than the fact that air temperature was close to the freezing point for most of the winter season. Hence, the differences in soil temperatures between the ski-slope and the natural snowpack outside the ski-slope were not significant according to the simulations, which was supported by the soil temperature measurements.

Chapter 6 is a study on snow hardness and snow stratigraphy with the classical hand and ram hardness methods and the SnowMicroPen method. Surface section photographs of snow samples from the investigated snow profiles serve as an objective reference. The hand test captured 80 % of all layers and layer boundaries and the ram test 60 %. The micro penetrometer captured the stratigraphy more complete than hand and ram profiles. The hand and ram profiles are practical but coarse methods to record the snowpack stratigraphy and snow hardness. Important details are missed, for example thin hard and soft layers which are highly relevant to avalanche formation. Differences in soft snow are resolved by the micro penetrometer, which is problematic or impossible with hand and ram tests. The hand and ram hardness tests indicate systematically greater snow hardness than the micro penetrometer. The surface sections and micro penetrometer profiles show a much more stratified snowpack than revealed in the classical snow profiles.

In conclusion, **Chapter 7** is a synopsis of the presented results and gives an outlook on future applications and on research in the systematic investigation of the variability of mountain snowpacks.

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Chapter 2

Developments in snow stratigraphy

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Abstract

Snow can be regarded as an aeolian sediment with rapidly changing properties in time and space. In this paper we explore the history of stratigraphy of snow, which follows intricate ways. The purpose is to show the different developments and then to develop a synthetic concept to optimally assess snow stratigraphy. The observation of snow stratigraphy starts in the 18th and 19th century with a geologic focus, descriptions are superficial and only verbal. In the early 20th century, the scientific interest in snow stratigraphy increases, detailed descriptions and drawings are available. This was also the time where larger scale geomorphologic features and surface processes were observed and documented. Starting from the 1940s, the interest shifts to the physical and mechanical properties in "homogenous" layers. In fact, we observe a paradigmatic change. The scale of interest shifts from a slope centred approach to a sample-centred approach. Stratigraphic description becomes one-dimensional, and the concept of well-defined layers gets accepted. This paradigm still predominates today. However, the formation of avalanches is inextricably related to spatial variation of mechanical properties. New instrumental developments show that the perceived strict layering may be a too simple model. Our study shows that a modern snow stratigraphy has to integrate different scales and requires a further development of fast and high-resolution modern instruments.

2.1 Introduction

The stratigraphy of snow is the framework for research of the properties, processes and dynamics of a natural snowpack. Unlike in soil science (USDA, 1993) and sedimentology (Selley, 1988), snow stratigraphy is not a discipline by itself. Snow stratigraphy is the total of the snowpack properties including the surface, the layers, the layer boundaries and the horizons without clear boundaries. It is the summary of the textural differences in a natural snowpack at all scales. The aim of snowpack stratigraphy is to capture the state of the snowpack at the appropriate resolution in order to derive the processes acting within the snowpack. Snow stratigraphy is used today as a description tool in avalanche warning, snow hydrology and snow research. The in-situ investigation of the snowpack is commonly achieved by digging and preparing a snow profile. The snow profile is a vertical section of

the snow cover, the state of the art is to record on a vertical line textural properties at a resolution of about 1 cm or more. The recent development of numerical models simulating the snowpack with layers (Bartelt and Lehning, 2002; Brun et al., 1989; Jordan, 1991) and of highly resolving instruments and sampling (Schneebeli, 2002) shows that the classical method is at its limit.
This study of the developments in snow stratigraphy follows the different directions that were taken historically in snow stratigraphy and it points to directions that yield a physically relevant representation of the complete snowpack. Snow science is manifested in a relatively small research community worldwide and the historical developments show a mix of consistent and scattered research directions. Some research directions are consistently followed, some are given up and some are readopted. First, we lay out the historical development in detail and summarise the main developments. Then we develop the requirements for a strategy of future snow stratigraphical work. We conclude on thoughts about the potential use of snow stratigraphy in snow mechanics and hydrology. A modern snow stratigraphy could contribute much to the understanding of snow processes and to the development of snow models.

2.2 Chronology of developments in snow stratigraphy

The chronological overview of snow stratigraphy in respect to the development of methods is given in Tab. 1. The chronology is divided into five time periods, which are prior to 1900, 1900 until the 1930s, 1940 until the 1960s, 1970 until the 1980s and 1990 until present. For each time period, the methodological, qualitative and quantitative developments are listed with the names of the first author of the contribution. The snow profile methodology is given in five groups. The descriptive method is the verbal interpretation of the stratigraphy. Optical methods are analog and digital optical recordings of the stratigraphy. Morphological methods capture the grain size and grain shape of snow crystals. Mechanical methods focus on the mechanical properties of the whole snowpack or layers. Textural methods serve to interpret the snowpack with combined morphological and mechanical properties. Following the table, the developments, major advancements and trends are then summarised for each period.

2.2.1 The 19th century and earlier

Descriptive methods

During the early 18th century the first observations of the phenomena snowpack and avalanches are documented by Scheuchzer (1706) in Switzerland. Snow stratification in particular is observed by Agassiz (1840), who quoted Zumstein and de Saussure as having known about it before. Tyndall (1861) observes and mentions stratigraphy of snow.
His interest is in glacier ice structure and its relation to the stratigraphy of snow and he reports on the observations made during numerous mountaineering expeditions in the Alps. The Swiss forester Coaz (1881) first recognises the relationship between the stratigraphy of snow and avalanche formation. The Swiss geologist Heim (1885), a pioneer in the field of glaciology, observes and physically explains the reasons for the stratified nature of the snowpack and of metamorphism. He also describes the changing optical properties and hardness properties during the metamorphism from new snow to glacial ice (firnification).

Table 1: Chronology of the developments in snow stratigraphy ("quant." is quantitative).

Method → Period ↓	Descriptive	Optical	Morphological	Mechanical	Textural
18th and 19th century	Observation and description of phenomena: Scheuchzer (1706), Agassiz (1840), Tyndall (1861), Coaz (1881), Heim (1885)				
1900 – 1930s	**Descriptive**	**Optical**	**Morphological**	**Mechanical**	**Textural**
	Observation and description snow texture and snow surface phenomena: Paulcke (1926,1938), Welzenbach (1930), Hess (1933), Seligman (1936)	Drawings and photographs of cross-sections: Paulcke (1926, 1938), Welzenbach (1930), Hess (1933), Seligman (1936), Eugster (1938) Dye method:Welzenbach (1930) Microscopy: Seligman (1936) Bonfire method: Nakaya (1936) Thin sections (quant.) : Bader et al. (1939)	Collection of grain shapes: Seligman (1936), Bader et al. (1939) Grain size, 3 grain size categories for field analysis, laboratory sieve analysis: Bader et al. (1939)	Density (quant), 2 hardness categories: Welzenbach (1930) Pressure gauge, shear frame: Eugster (1938) Finger hardness test: Paulcke (1938) Hand hardness test, 4 hardness categories, ramsonde, hardness related to failure planes: Bader et al. (1939)	Snow surface categories: Hess (1933) Textural classification, snow surface categories: Paulcke (1938) Combined morphological and mechanical snow classification: Bader et al. (1939) Textural analysis: Bader et al. (1939)
1940 – 1960s	**Descriptive**	**Optical**	**Morphological**	**Mechanical**	**Textural**
		Thin section analysis (quant.): De Quervain (1948) Translucent profile: De Quervain (1948) Translucent profile advanced: EISLF (1951) Translucent profile advanced: Andersen (1960) Translucent profile (quant.): Benson (1962)	Measured grain size (quant.): EISLF (1951) Grain shape classification: EISLF (1951) Snow classification (quant.): Eugster (1952) First Int. snow classification: Schaefer et al. (1954) Field guide to snow crystals: LaChapelle (1969)	Safety index: Bucher (1948) Measured tensile and compressive strength, cone penetrometer, hand hardness scale (fist-knife), relation to ramhardness(quant): De Quervain (1948) Parallel hand hardness and ram hardness: EISLF (1951) Hardness related to failure planes: Atwater et al.(1953)	Relation of hardness and grain size with tensile strength (quant.): De Quervain (1948) Textural analysis (quant.): Eugster (1952) Snow surface classification: CRREL (1962)

1940 – 1960s continued	**Descriptive**	**Optical**	**Morphological**	**Mechanical**	**Textural**
		Relation of spectral extinction to grain size and density (quant.): Mellor (1966)		Snow resistograph: Bradley (1966) Shear, tensile and compressive strength, shear frame (quant.): Roch (1966)	
1970 – 1980s	**Descriptive**	**Optical**	**Morphological**	**Mechanical**	**Textural**
	Description of mountain forest snow cover: In der Gand (1978)	Thin section analysis (quant.): Good (1974) Dye method : Kovacs (1976) Relation of radar backscatter to water content and density (quant.): Ellerbruch et al. (1977), Boyne et al. (1979,1980) Drawings of mountain forest snow cover : In der Gand (1978) Scanning photometer, low scan video (quant.): Good (1980) Thin section analysis (quant.) : Good (1982) Thick section sampler for whole profile : Harrison (1982) Photography of cross-sections of mountain forest snow cover: Imbeck (1983,1987) Relation of FMCW radar to water equivalent (quant.): Gubler et al. (1984)	Near surface faceting: Armstrong (1977) Extension of snow classification by the degree of metamorphism and riming: Ferguson (1984, 1985)	Improvement of ram hardness equation (quant.): Gubler (1975) Higher resolution density and temperature measurements (quant.): Ferguson (1984, 1985) Digital resistograph (quant.): Dowd et al. (1986) Cone penetrometer (quant.): Schaap et al. (1987) Rutschblock test: Föhn (1987a,b) Stability classification (quant.) : De Quervain et al. (1987)	Textural analysis (quant.): Keeler (1969), St. Lawrence (1974), Kry (1975a,b) Textural parameters from thin sections: Good (1974) Textural analysis (quant.): Gubler (1978) Relation of distance between grains with tensile strength (quant.): Good (1987) Classification for mountain forest snow cover: Imbeck (1983, 1987)

1990 - present	Descriptive	Optical	Morphological	Mechanical	Textural
		FMCW radar in mountain forest snowpack: Gubler et al. (1991) Translucent profile method, improved: Good et al. (1992) Surface sections vertical and horizontal: Good et al. (1992) FMCW radar analysis (quant.) : Koh (1993) Dye method improved : Kovacs (1993) Surface sections (3D) : Schneebeli (2000) Nuclear magnetic resonance : Ozeki et al. (2000) X-ray micro tomography: Coleou et al. (2001), Schneebeli, (2002) NIR photography (quant.): Haddon et al. (1997), Schneebeli (2002)	Int. Classification for seasonal snow on the ground: Colbeck et al., 1990) Snowpack model SAFRAN CROCUS MEPRA, dendricity, sphericity (quant.): Brun et al. (1992), Lesaffre et al. (1998) Near surface faceting: Fukuzawa et al. (1993) SnowPit, spatial or temporal snow profile: Shultz et al. (1998) Near surface faceting (1mm layers): Birkeland et al. (1998)	Digital resistograph improved (quant.): Brown et al. (1990) Weak layer analysis with Rutschblock test: Föhn (1992) Loaded column test (quant.): McClung et al. (1993) PANDA penetrometer (quant.): Navarre et al. (1994) Rammrutsch test, Schweizer et al. (1995) Compression test (tap test): CAA/NRCC (1995) Stuffblock test: Birkeland et al. (1996) Snow micro penetrometer (quant.): Schneebeli et al. (1998,1999) Stability classification, advanced: Stoffel et al. (1998), Schweizer et al. (2001) Quantified loaded column test (quant.): Landry et al. (2001) Comparison of hardness methods (quant.): Pielmeier et al. (2002) Snow penetrometer (Sabre): Mackenzie et al. (2002)	First surface sections of heterogeneous snow samples and layer boundaries (quant.): Good et al. (1992) Relation of image patterns with mechanical properties (quant.): Good et al. (1992) Climatic snow cover classifi-cation : Sturm et al. (1995) Texture analysis (quant.): Schneebeli et al. (1999) Micro structural and mechanical model (quant.): Johnson et al. (1999) Relation of NIR to stratigraphy: Haddon et al. (1997), relation to grain size (quant.): Schneebeli (2002)

Summary of period 19th century and earlier

The earliest documentations of snow stratigraphy are purely descriptive. No documentations of quantitative snow stratigraphic investigations exist from this period. During this time, the close observation of the phenomena “snow” during expeditions to Greenland, polar and alpine regions begins. Large time and spatial scales are covered by these observations and led to a geomorphic-sedimentologic perception of the snow cover. For the first time systematic methods for snow investigations and avalanche observations are developed.

2.2.2 From 1900 to the 1930s

Descriptive methods

The first systematic snow profile documentations in the Alps are recorded by Paulcke (1926; 1938), Welzenbach (1930) and Hess (1933). They observe and describe snow stratigraphy, snow texture and snow surface phenomena in terms of grain size, hardness and humidity and also document the first systematic methods and measurements.

Optical methods

Dye method:

Welzenbach (1930) uses dye tracer experiments to visualise the stratigraphy of snow and its effect on water transport in the different layers (Fig. 1). Another dye method for this purpose is introduced by Nakaya, Tada et al. (1936), the so-called bonfire method. It is a simple dye-method where smoke is used to stain the snow profile for visualization of the snow stratigraphy. The method produces high contrast for black and white photography of snow stratigraphy. Solar and wind exposure, melting and considerable effort for opening the snow pit are disadvantages of this method.

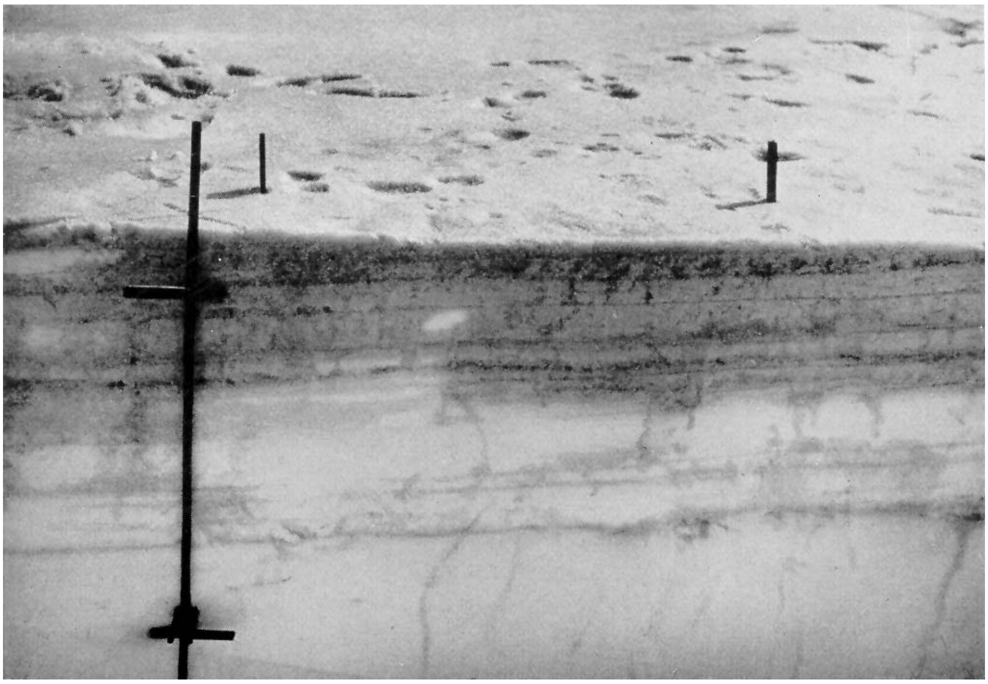


Fig. 1: Snow stratigraphy proven by a dye tracer experiment in 1930 (Welzenbach).

Drawings, photography and microscopy:
Welzenbach (1930) investigates the development of cornices and uses the method of two-dimensional snowpack cross sections to document different types of cornices (Fig. 2).

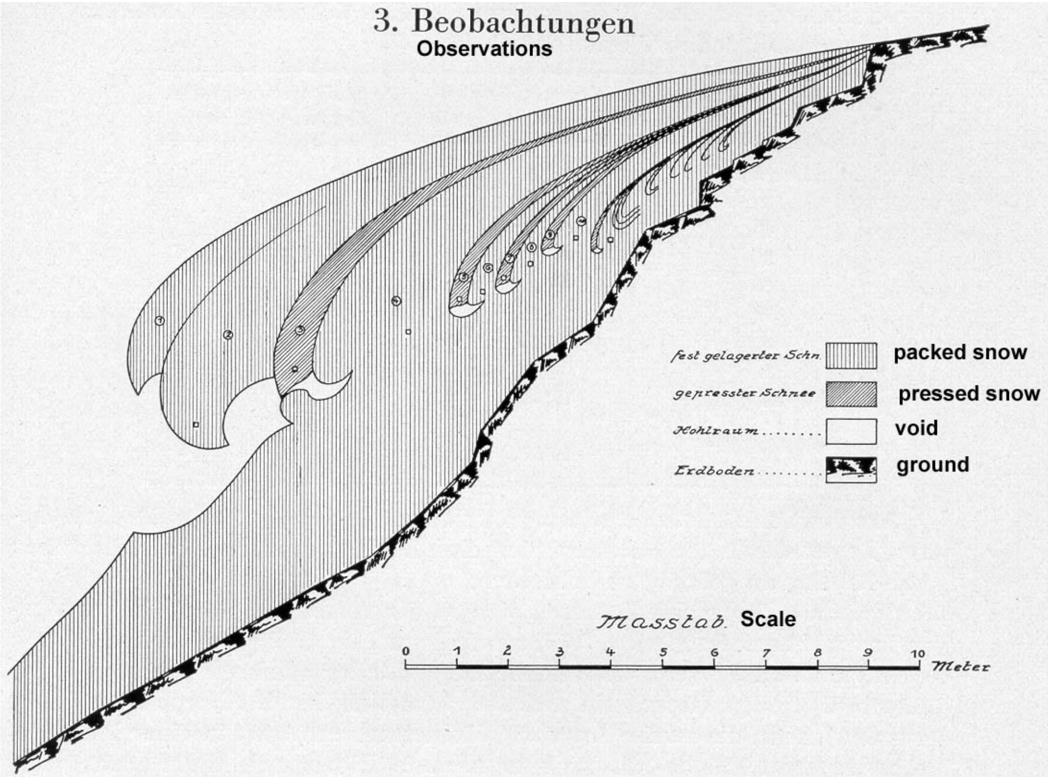
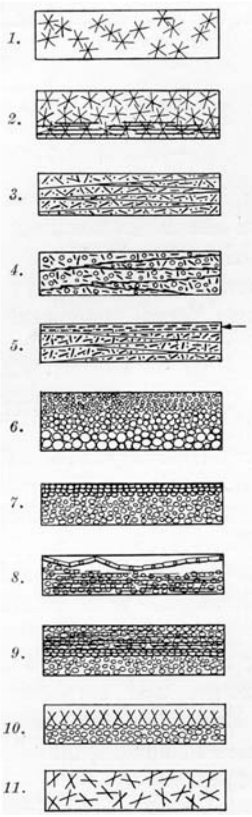


Fig. 2: Two-dimensional snow profile of a cornice drawn by Welzenbach (1930). The stratigraphy (layers 1-12) was established by density measurements.

Paulcke (1926), a geologist, started to work on the fundamentals of snow and avalanche science 40 years prior to his comprehensive publication on practical snow and avalanche science (1938). He deals with both, snow classification and stratigraphy and applies geological methods especially to the snow stratigraphy. Paulcke classifies the snowpack in terms of sedimented layers which may contain snow with homogeneous but also with heterogeneous properties. He also considers the geomorphological settings of the snow cover. His drawings of snow profiles in the form of cross-sections are very detailed and illustrative. Examples of snow affected by wind perturbations and snow with gradually changing grain size are shown in Fig. 3. His snow classification is based on a two-dimensional approach to discrete layers and horizons with continuously changing properties. According to Paulcke (1938) it is necessary to picture the snowpack on a slope as cross section (Fig. 4) to judge the avalanche danger. Hess (1933) shares this opinion and speaks of thin separation planes between snow layers in the snowpack, and of snow stratigraphy as a suitable method to record them. He documents detailed graphical profiles and photos of pit walls and shows dividing planes as thin as 1 cm in dry and wet snow, which were responsible for the formation of avalanches (Hess, 1933). Seligman (1932-1934) translates and expands Welzenbach's work on snow deposits. He gives detailed description of his stratigraphic methods, which were photography (cross sections of field profiles) and binocular microscopy (snow crystals). He observes and describes processes such as firnification, snow surface formation and avalanche formation based on his photographic and microscopic documents.



1. “wild snow - new snow”

2. “dry and
wet new snow”

3. “packed snow with cross lamination”

4. “pressed snow with cross lamination”

5. “wind pressed packed snow, wind crust
overlaying packed snow with cross lamination”

6. “fine, medium
and coarse grained firn”

7. “wind crust - harsch
overlaying firn, dry“

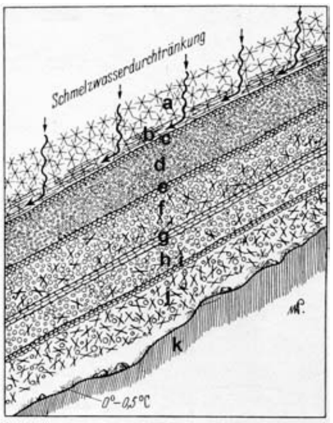
8. “sun crust - firn spiegel
firn, wet“

9. “slack flow, ice lamella built from harsch layer
overlaying firn”

10. “surface hoar
overlaying firn”

11. “depth hoar”

Fig. 3: Paulcke’s (1938) descriptive snow classification



a) “loose snow – new snow”
b) “slack flow, hydroplane on c) wind crust = avalanche danger!”
d) “firn, fine grained”
e) “wind crust – harsch”
f) “firn, medium grained, with singular facets”
g) “ice lamella from sun crust”
h) “firn with singular facets”
i) “crust”
j) “depth hoar with singular firn grains, development of voids = avalanche
danger!”
k) “bedrock”

Fig. 4: Snow profile in the form a descriptive cross section of the snowpack (Paulcke, 1938)

In his comprehensive monograph “Snow Structure and Ski fields”, Seligman (1936) provides a broad collection of all ice forms. He also gives a comprehensive overview of available instruments and methods in snow research (photography, microscopy, penetrometry by a pressure gauge, psychometry, and anemometry). His work is the basis for further snow research where the focus starts to shift from a geomorphologic-sedimentologic perception of snow stratigraphy to the study of the physical properties of homogeneous snow samples. Seligman (1936) qualitatively compares section photographs (Fig 5), to his drawings of cross sections of slope profiles (Fig. 6). From his empirical knowledge and snow stratigraphic experience he draws conclusions about snowpack processes such as avalanche formation and metamorphism.

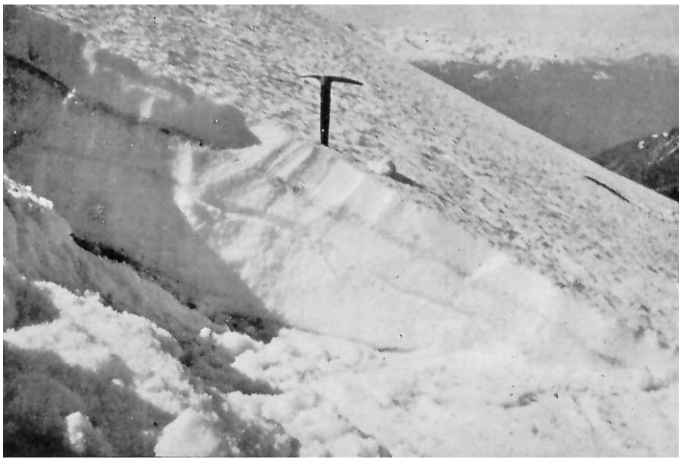


Fig. 5: “A snow section where two crust layers betweeen the three snow strata are clearly seen (see Fig. 6, A). If the ice layers are frozen to the snow strata, there is no danger of the snow sliding off by strata.”: (Seligman, 1936, p. 310)

Thin section:

To preserve the texture of snow aggregates the method of thin sections is developed and a quantified, micro textural analysis based on thin section images is introduced (Bader et al., 1939).

Morphological methods

From 1934 to 1938, Bader, Haefeli, Bucher, Thams, Neher and Eckel (1939) carry out their systematic work on snow and avalanche research. One of their goals is to find methods to characterise snow crystals in order to record snow profiles. For this task Bader (Bader et al., 1939) uses a mineralogical and crystallographic approach to homogeneous snow. In laboratory experiments, Bader et al. (1939) focus on the detailed measurement and classification of homogeneous snow samples and isolated snow crystals using micro photography and microscopy. They provide a comprehensive collection of grain photographs and thin sections and lay the foundation for the analysis of snow micro structure (Bader et al., 1939). They also establish three grain size categories for snow profiles taken in the field, which are fine, medium, coarse.

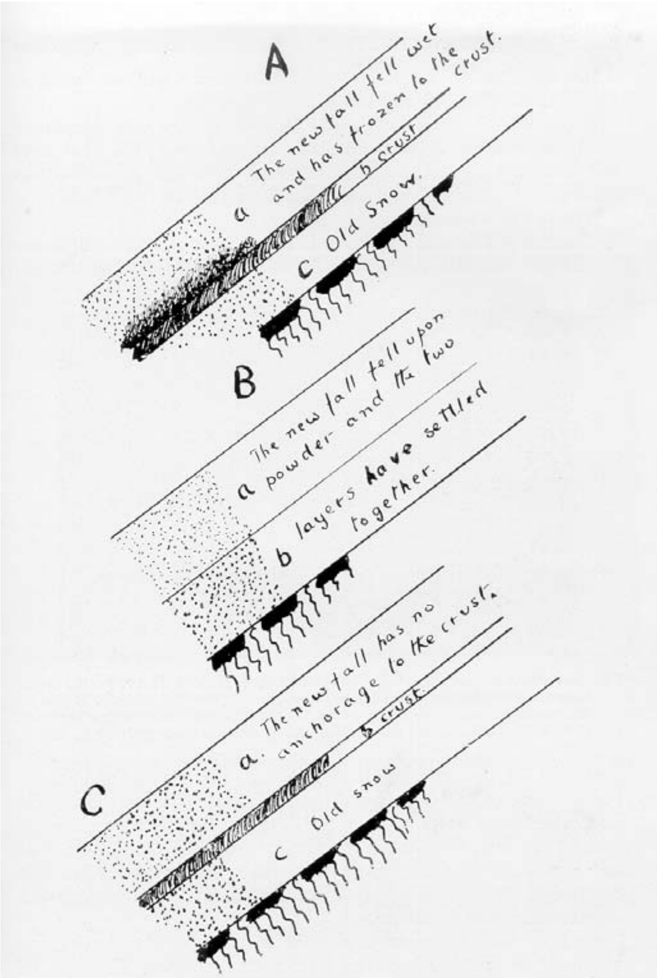


Fig. 6: "In A the NEW SNOW *a* fell wet and froze on to the CRUST LAYER *b*. It is therefore well anchored to it. A photograph of this state of affairs is shown in Fig. 5. In B the NEW FALL *a* fell upon POWDER SNOW *b* and the two have settled together. In C the NEW FALL *a* fell at low temperatures and has no anchorage to the CRUST LAYER *b* on which it lies. (Seligman, 1936, p. 309)

Mechanical methods

Density, hardness and strength measurements:

Welzenbach documents density profiles of cornices. Tab. 2 is the density profile of the cornice cross-section shown in Fig. 2. These are the first quantitative records of snow profiles, which are motivated by research on cornice formation, water flow in the snowpack and slab avalanche formation. Welzenbach (1930) qualitatively distinguishes two hardness categories from textural differences observed by eye that are packed snow and pressed snow.

Paulcke's (1938) technique of snow profiling is to visually inspect the open profile wall before sensing the hardness differences by hand. The hand hardness is described in terms of ease at which a layer can be penetrated by a finger. This is the first description of a hand hardness test. Paulcke also measured density and grain properties.

The focus of Eugster's (1938) snow studies is to obtain a more profound understanding of the snowpack for the planning and construction of expensive permanent avalanche protection

Table 2: First quantitative snow profile classified by hardness (degree of packing) and measured densities of the layers in the cornice shown in Fig. 1 (Welzenbach, 1930, p. 16).

Layer No.	Specific weight in moderately dense layers (packed snow)	Specific weight in dense layers (pressed snow)
1 and 2	0.220	
3		0.420
4	0.352	
5		0.466
6	0.400	
7		0.480
8	0.425	
9		0.490
10	0.440	
11		0.500
12	0.450	

devices. He recognises the importance of thin layers and the limitations of the stratigraphic methods available, for example the low-resolution density profile. First, Eugster develops a new instrument, the shear frame, to measure the cohesive strength between selected layers. Then he develops a pressure gauge, an instrument to measure the penetrability of snow (Eugster, 1938). This instrument delivers a quantitative profile of the cohesive strength of the snowpack. The measured pressure during penetration is also an indication of the cohesion between layers in a snowpack (Seligman, 1936). Eugster measures the cohesive strength of snow throughout the winter 1935 in shady and sunny slopes and documents the strength profiles. Fig. 7 shows two examples of quantified and graphically documented snow profiles, including stratigraphy, morphology and mechanical strength. These snow profiles are amongst the first quantitative mechanical profiles that relate mechanical, stratigraphic and morphologic properties. The way the morphology is drawn shows that gradually changing snow properties exist. It also shows that there is morphological and textural variation within the stratigraphic layer.

Haefeli's (Bader et al., 1939) snow mechanical goal is to determine the properties that make a snow layer a potential avalanche active layer. They establish an ordinal snow hand hardness scale, which consists of four classes: loose, soft, hard, very hard. In order to objectively measure the mechanical snow hardness, Haefeli developed the Swiss ramsonde from penetrometers used in soil mechanics (Fig. 8). It has a relatively large measuring cone (diameter 40 mm) and a 60° included angle. Because of its low spatial resolution Haefeli (Bader et al., 1939) cautions when interpreting the ram hardness profile. Thin and soft layers often responsible for the formation of avalanches can not be resolved with the ramsonde. Bader et al. (1939) also describe the technique of snow profiles at the level Weissfluhjoch study plot and suggest that the classical snow profile and mechanical ram profile can provide some characterisation of the physical snow properties, but they are not suited to record the data at the necessary spatial resolution. The advantage of in-situ tests is obvious though. The handling of snow samples for laboratory analysis is difficult, densification and subtle or obvious failures often occur in soft or brittle snow samples (Bader et al., 1939).

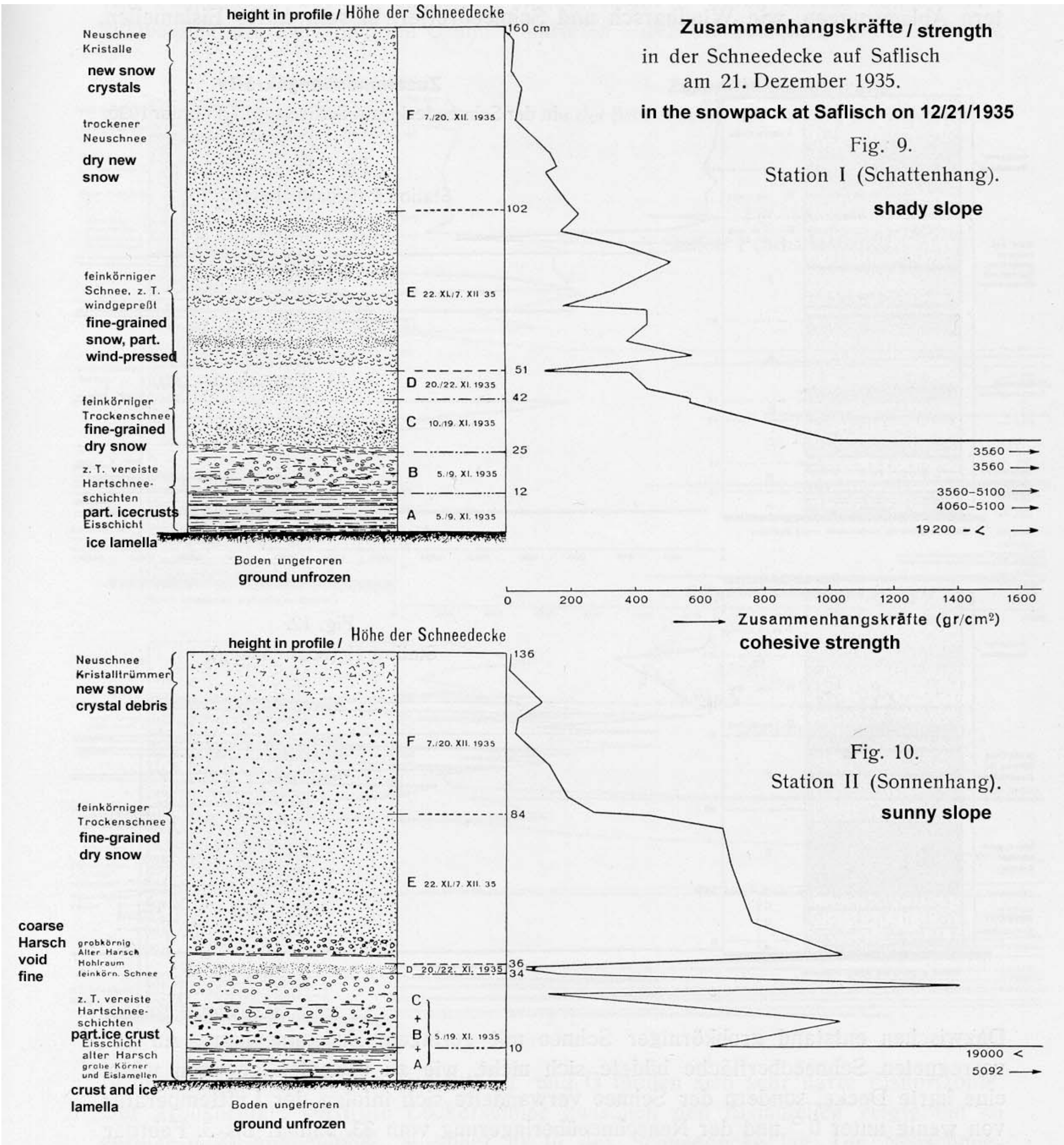


Fig. 7: Two snow profiles in terms of morphology and cohesive strength by Eugster (1938).

Textural methods

In his illustrative snow classification, Paulcke (1938) classifies the snow within homogeneous and heterogeneous layers according to grain size and grain morphology as well as according to the degree and pattern of packing, pressing and melting. This is the first time that a textural approach is taken in a snow classification. Paulcke also describes snow surface categories. Bader et al. (1939) give a detailed description of snow profile recordings at the Weissfluhjoch level study plot where the stratigraphy is established by placing a thread on the snowpack after each new snowfall. This marks the layer boundaries when the snow pit is opened and the height of each layer can be measured between the threads. Temperature is measured in 10 cm intervals. Each layer is sampled by a cylinder and density, water equivalent and vertical air permeability are determined for each layer between threads.

Fig. 8: Swiss Ramsonde was developed by Haefeli (Bader, Haefeli et al., 1939) to obtain an objective mechanical hardness profile of the snowpack.

An example of a complex, quantified study plot profile is shown in Fig. 9a and a complex slope profile where ram hardness is related to an avalanche failure plane is shown in Fig. 9b. Fig. 10 shows a unique textural snow classification system developed by Bader et al. (1939). It combines ordinal grain size and hardness classes (loose, soft, hard, very hard). Bader et al. (1939) illustrate the first time profiles, which show the development of the snowpack over a whole winter season. It is accomplished by graphically interpolating a series of bi-weekly snow stratigraphic profiles (Fig. 11). The display of the simultaneous meteorological parameters is included. A quantified micro textural analysis from the thin section images is introduced by Bader et al. (1939), where principle axes are measured in the thin sections before and after deformation experiments. The samples are exposed to compressive, shear and tensile deformation experiments and the changes in snow micro texture are documented.

Summary of period 1900-1930s

In the beginnings of the 20th century different professionals, the Army and mountaineers begin systematic investigations of the seasonal snowpack. The observations and experiments are carried out during numerous expeditions in the first three decades of the 20th century. Thereafter, prominent extensive contributions are published in the 1930s. During this period, snow morphological, mechanical textural properties become quantified and snow surface classifications are established. Special snow instruments and also analytical methods are developed to gain objective and quantitative data. There is a call for objective methods with higher resolution than the one’s available at the moment. A shift from a snowpack perception in its geomorphologic-sedimentologic settings to a focus on the physical properties of snow takes place towards the end of this period.

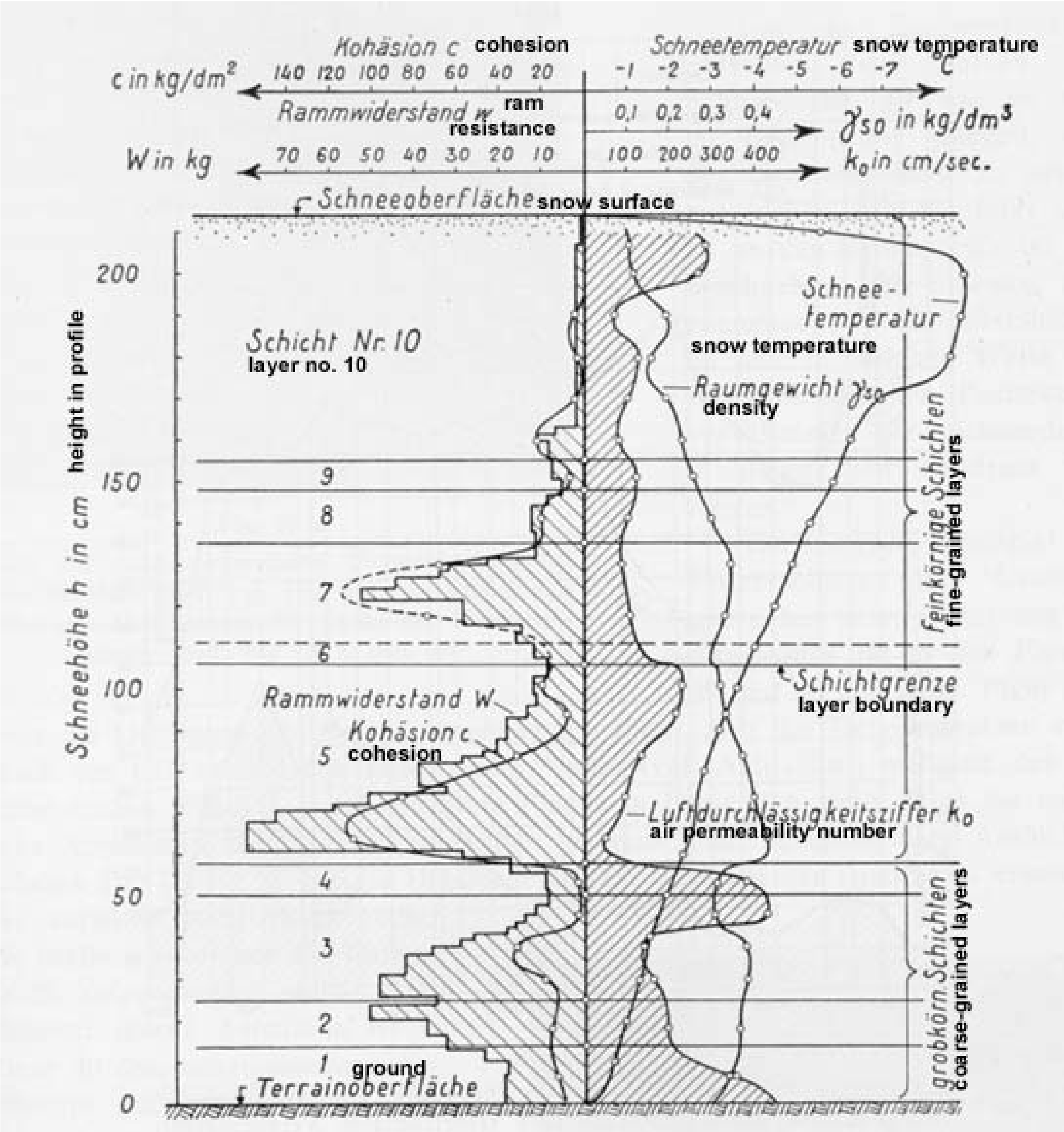


Fig. 9a: Complex snow profile taken by Haefeli on 16 February 1937 at the flat study plot Weissfluhjoch, Davos (Bader et al., 1939). Measured parameters are the stratigraphy based on the threads, ram resistance, cohesion, air permeability, density, snow temperature and grain size categories.

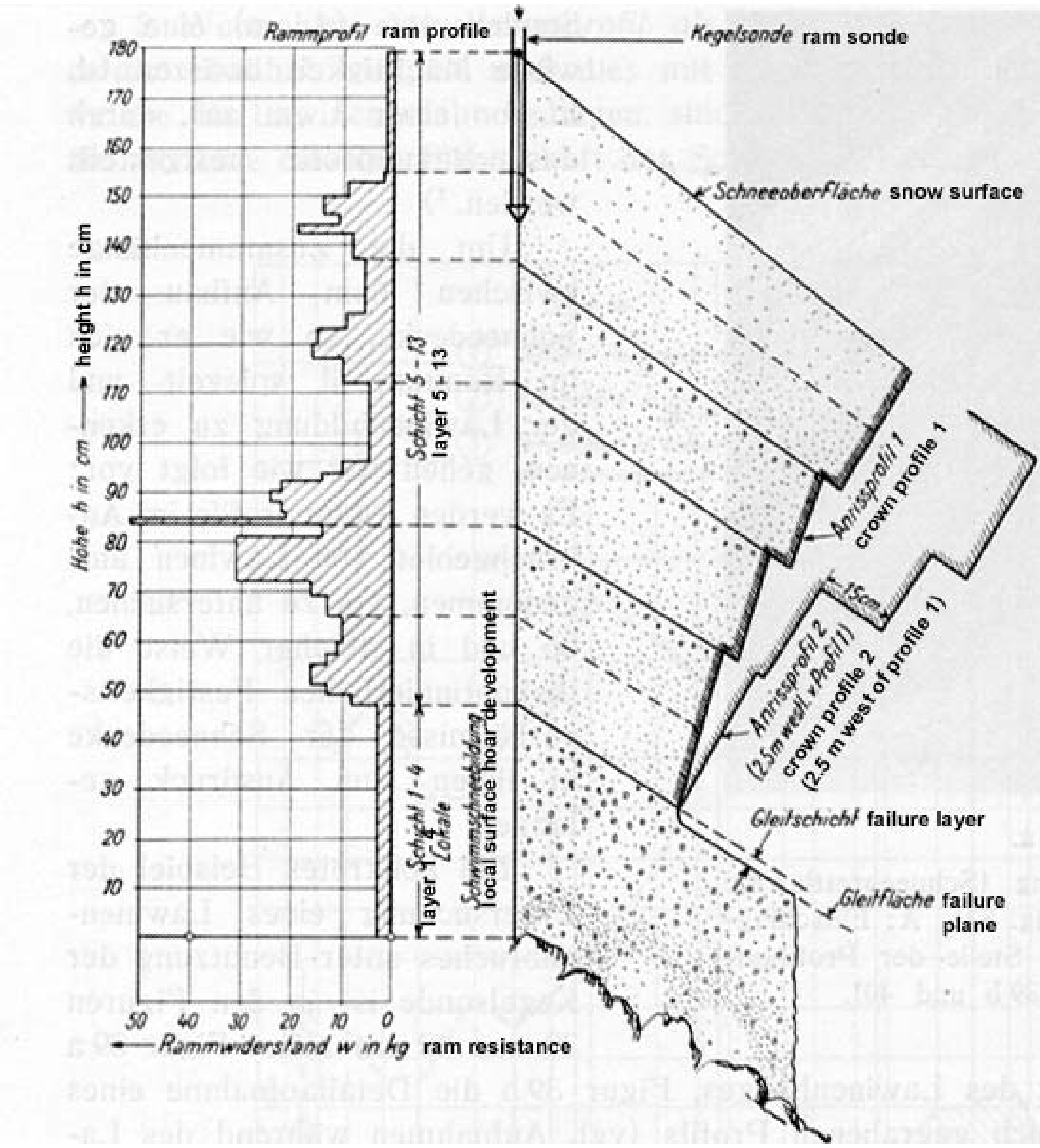
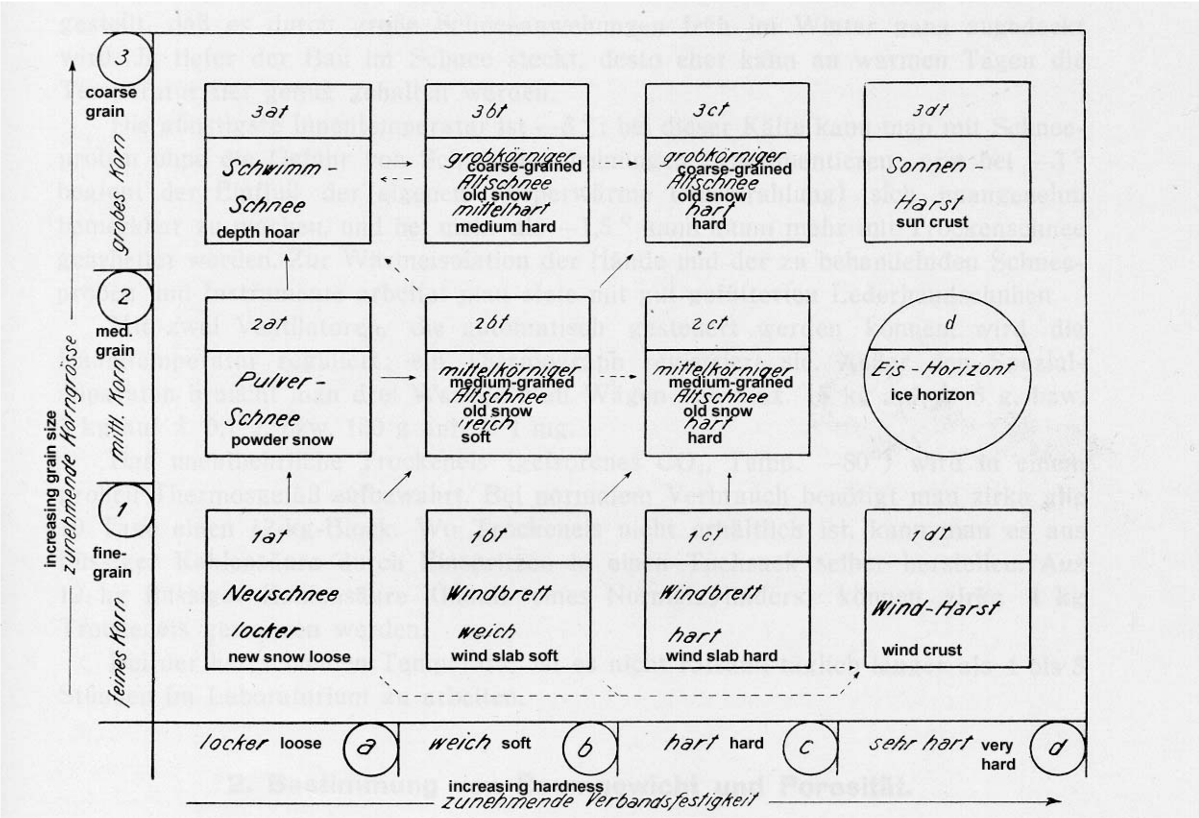


Fig. 9b: Complex slope profile taken by Haefeli above an avalanche crown (Bader et al., 1939). Measured parameters are the stratigraphy based hardness differences, ram resistance. Grain size and morphology are relative because they are only drawn.



Definitions:	1	fine grain	a	loose or soft	n	wet
	2	medium grain	b	medium hard	t	dry
	3	coarse grain	c	hard		
			d	very hard		

Fig. 10: Snow classification system field profiles by Bader et al. (1939). The classification combines categories of grain size (visual observation) and hand hardness (manual observation). Wetness and dryness are also indexed.

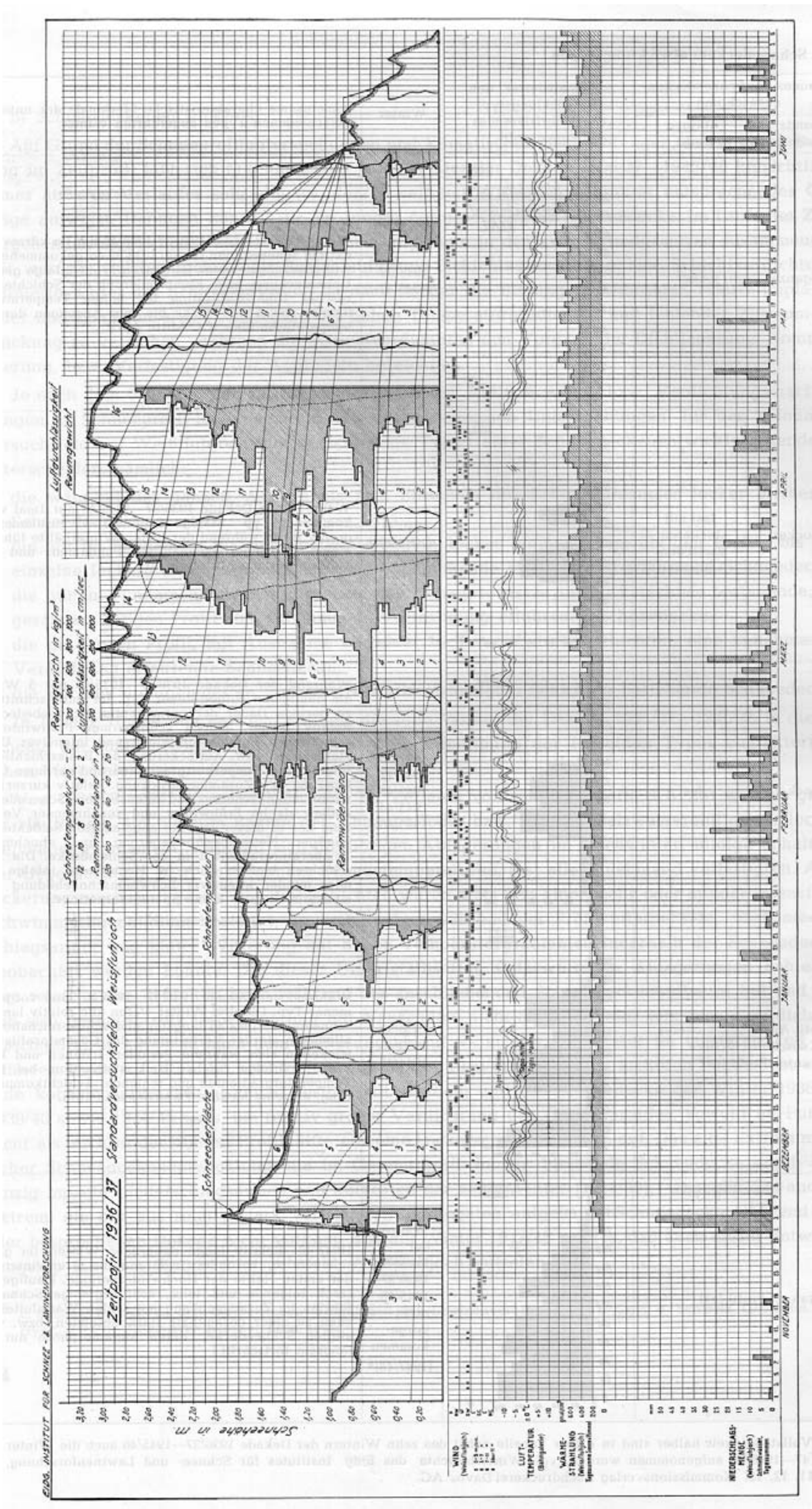


Fig. 11: Complex time profile taken by Bader et al. (1939) at the study plot Weissfluhjoch, Davos during the winter 1936/37. In the upper part monthly snow profiles, in the lower part daily meteorological parameters are drawn.

2.2.3 From 1940 to the 1960s

Optical methods

Thin sections:

De Quervain (1948) uses thin sections of snow samples to analyse snow texture that is the arrangement of grains in the snow aggregate.

Translucent profiles:

De Quervain (1950) documents the first translucent snow profile, which he correlates to classical snow stratigraphy and a number of different snow hardness measurements (Fig. 12).

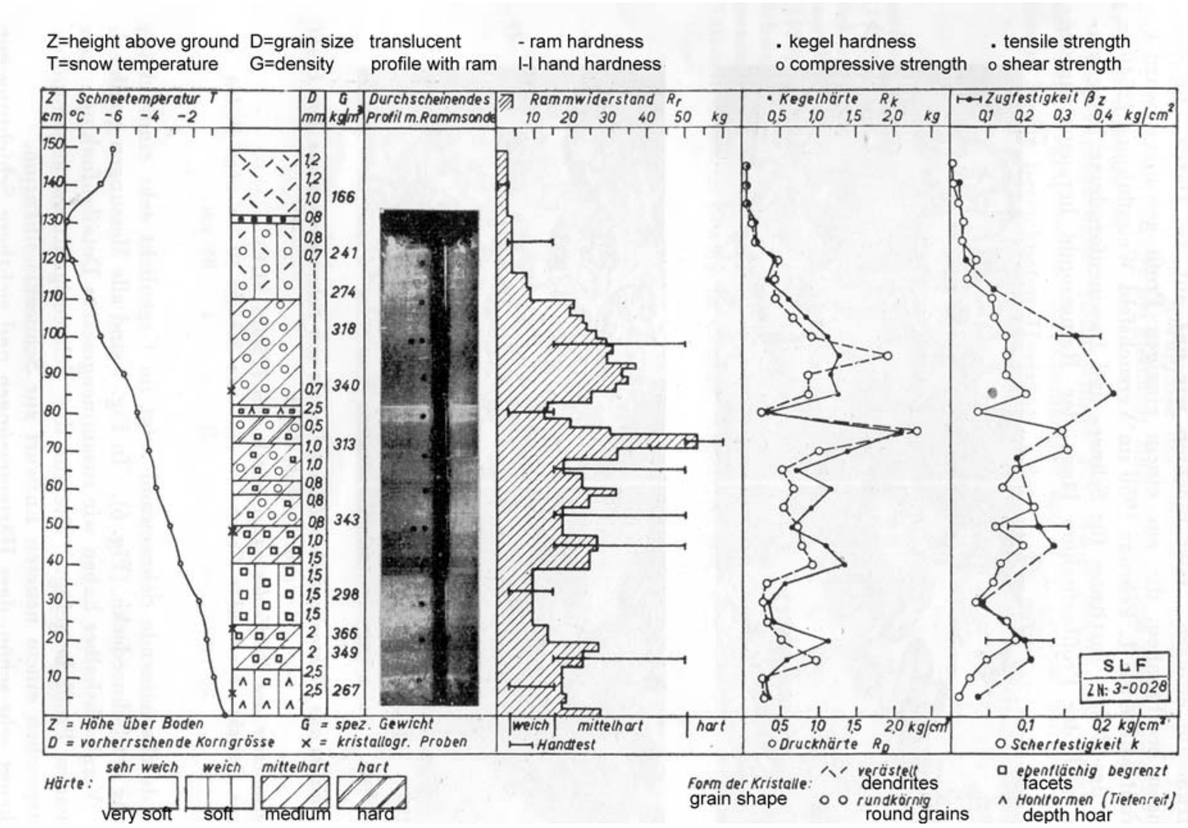


Fig. 12: Snow profile taken on 1 February 1950 with a translucent profile and a comparison of hardness measurements (De Quervain, 1950).

Translucent profiles appear in documentations of qualitative analysis of snow profiles taken at avalanche fracture crowns (EISLF, 1951). Andersen (1960) develops and documents a photographic technique for recording snow stratigraphy from translucent profiles. To emphasise hardness discontinuities, he brushes the profile parallel to the stratification with a whiskbroom. Andersen states that the layer resolution is improved to 1 cm with this method. Benson (1962) analyses the spatial variability of the snowpack on the snow pit scale (1 m). He finds that discontinuous wind slabs and ice lenses exist which can falsify the interpretation of the stratigraphy of the annual accumulation on inland ice. He uses pit wall and thin section photography and shows a correlation of the strata of high and low density with dark and light strata on the photos (Fig. 13). His translucent profiles illustrate the great vertical variation of the natural snowpack.

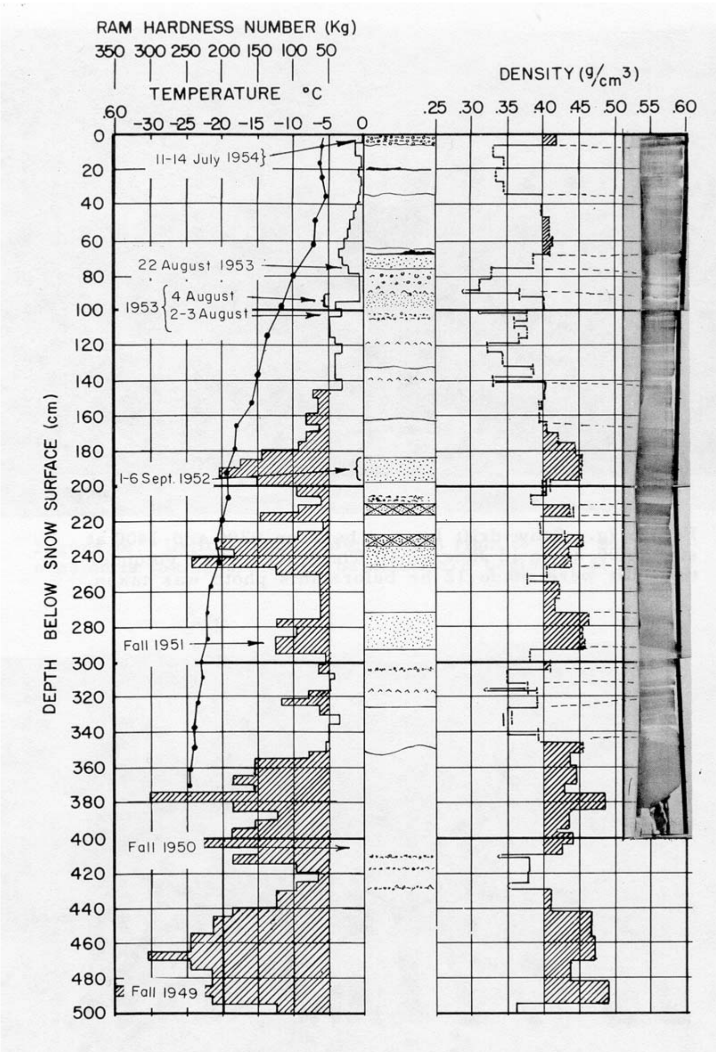


Fig. 13: Vertical snow stratigraphy and photo section (right). Strata of high and low density correlate with dark and light strata respectively on the photos. Layer boundaries are not always horizontal (Benson, 1969).

Spectral extinction:
Studies on optical properties of snow and their relationship to grain size and density are presented by Mellor (1966). The spectral extinction in snow is applied to the measurements of snow structure and for the remote sensing of snow-covered terrain.

Morphological methods
Measured grain size and classified grain shape are included in the snow profile procedure at the study plot Weissfluhjoch, Davos (EISLF, 1951). Because grain diversity plays such an important role, Eugster presents a new morphological classification that takes into account the actual variation of the grain shapes within one stratigraphic layer. Fig. 14 shows a snow profile where a distribution of grain shapes in quantiles of tenths is given for each layer (Eugster, 1952). However, this advanced snow morphological stratigraphy was not followed until today. The first International snow classification for solid precipitation and deposited

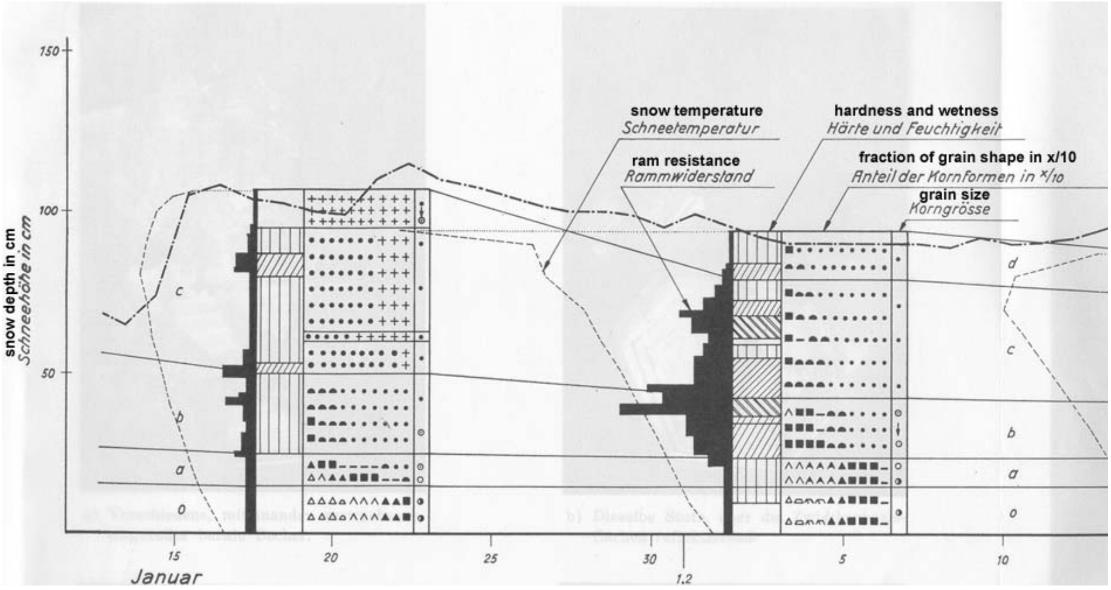


Fig. 14 Vertical snow profile section according to Eugster's (1952) expanded morphological classification where fractions (in tenths) of grain shapes are given for each layer.

snow (Schaefer et al., 1954) is a morphological classification for solid precipitation as well as the first standard method to measure and describe snow on the ground. It defines standard units and classes for rational and ordinal parameters measured and observed in the seasonal snowpack. The ordinal classes for wetness, grain shape and hand hardness are refined as compared to the previously applied schemes. Temperature and density profiles are taken in 10 cm intervals from the snow surface to the ground. International symbols for grain shape, grain size wetness and surface conditions are introduced. Fig.15 is an example snow profile taken according to the new international standard (Schaefer et al., 1954). This standard is a one-dimensional snow profile with discrete layers consisting of homogeneous snow properties and surface parallel layer boundaries. The temperature profile is drawn continuously with interpolated values between the 10 cm interval measurements.

LaChapelle (1969) presents a field guide to snow crystals, which includes an illustrative and abundant photographic collection of crystal types: He also gives a description of his photographic technique for snow crystals. Furthermore, LaChapelle (1969) shows and discusses the main systems of snow classification available: after Nakaya (1954), after Magono and Lee (1966), after Sommerfeld (1970) and after the International Snow Classification (Schaefer et al., 1954). Some snow classifications pertain to falling snow crystals others to crystals on the ground or a combination thereof, but all snow classifications exclusively deal with snow crystals and their individual properties.

Mechanical methods

Hardness and strength measurements:

Weak layers and their relationship to avalanche formation become a main topic in snow research. Bucher (1948) develops a safety index from laboratory shear experiments. The safety index is the relation of the stresses in snow to the snow stability and gives information about the snow stability. There is a need for mechanical measurements with higher vertical resolution than the ramsonde, in order to directly assess snowpack stability in the field (EISLF, 1950). De Quervain (1948) analyses snow texture, the arrangement of grains in the

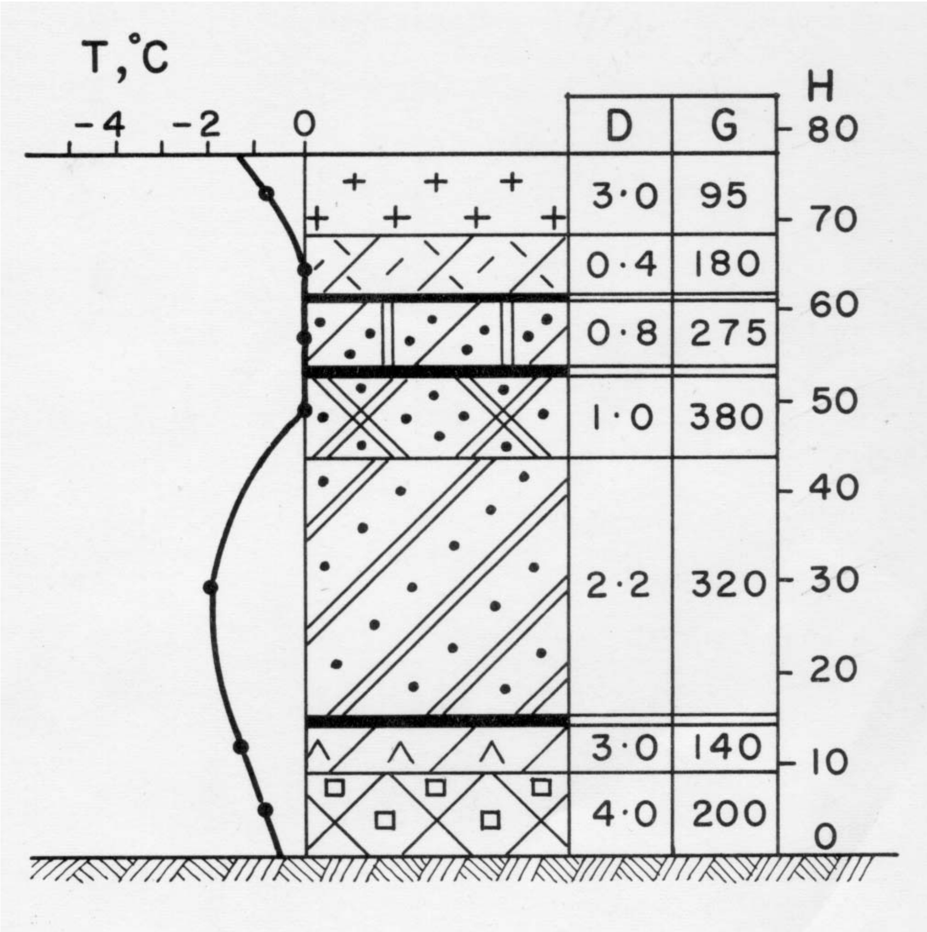


Fig. 15: Vertical snow profile section according to the International Snow Classification (Schaefer, Klein et al., 1954)

snow aggregate, by determining the spatial orientation of the grains in thin sections of snow samples. However, he is not able to establish a good relationship of the texture data and the snow mechanical properties with the methods available (ram hardness, tensile and compressive strength). He proposes to continue to focus on grain shape as an important factor for snow strength. De Quervain (1950) develops the kegelsonde, a small hand-held penetrometer with a cone diameter of 10 mm and a cone angle of 60°. De Quervain compares three different penetrometers, the ramsonde (vertical measurement), the kegelsonde and the pressure gauge (Eugster, 1938) (horizontal measurements). The correlation of the resulting profiles to the absolute scale determined from tensile and shear strength of snow is not satisfying and only crude comparisons can be made (Fig. 12). De Quervain finds that at equal ram hardness coarse grained snow has a lower tensile strength than fine grained snow. Hence, a better correlation can be achieved when grain size is included into the analysis. The parallel translucent profile photography supports this proposition. It is the first systematic stratigraphic and translucent profile documentation. At the same time, De Quervain (1950) proposes the use of an extended, device-independent hand hardness test, which is similar to mineralogical tests. He suggests five hand hardness classes that are correlated to classes of ram hardness, shown in Tab. 3. This hand hardness system, slightly modified, is still used today.

Table 3: Hand hardness method, classes and correlation to ram hardness by De Quervain (1948)

Device	**Fist** (with glove)	**Hand** (4 straight fingers, with glove	**Finger** (1 straight finger)	**Pencil** (sharpened)	**Knife blade**
Hardness class	Very soft	Soft	Medium Hard	Hard	Very hard
Ram resistance (kg)	0-3	3-15	15-50	50-150	> 150

Atwater and Koziol (1953) publish the first U.S. avalanche handbook where cone penetrometer tests in slopes that were subsequently artificially triggered are documented. The failure planes are depicted in the ram hardness profiles and related to large hardness discontinuities in the snowpack. Bradley (1966) invents a new snow mechanical instrument, the snow resistograph, to improve the shortcomings of the ramsonde. The snow resistograph uses an upward moving snow blade to record the resistance and gives a graphical output of the hardness profile right after the field measurement. The angle of the blade is about 10°, the surface area is 6.7 cm^2. Bradley (1966) states that the resistograph allows stability assessments of the snowpack. It did not become widely used, perhaps its weight and cost prevented it from becoming accepted as a new instrument for objective snow stratigraphic applications. Roch (1966) documents shear frame tests to study the tensile strength of snow. He analyses the relationship of the tensile strength of snow with varying temperature, time, overload and metamorphism.

Textural methods

The snowpack observations at the study plot Weissfluhjoch in Switzerland have a stratigraphic resolution of 1 to 2 cm. The classification that combines categories of visual grain size and hand hardness as well as classified grain shape is included in the bi-weekly stratigraphic profiles. The time profiles are combined with the available avalanche information, which supports an integrated analysis of snow depth, snowpack properties and avalanche activity. Starting with the winter 1943/44, the threads are no longer placed onto the study plot snowpack after every major snow fall, but the regular interval of each bi-weekly snow profile observation date. This causes a loss of the stratigraphic information about layer boundaries between snow fall events.

Eugster (1952) applies a quantitative texture analysis method to snow that was originally designed for the textural analysis of geological bodies (Sander, 1950). Eugster shows that the relative bond diameter (“Kornbindungsmass”) is correlated with the tensile strength of snow. This approach combines snow mechanical behaviour and snow textural properties. Eugster attributes the scatter of the data to his assumption of idealised spheres in his theoretical model underlying his textural parameters. This analytical snow textural approach was later readopted by Keeler (1969) and Kry (1975a; 1975b).

In 1962 instructions for making and recording snow observations are issued in the United States (CRREL, 1962). They provide guidelines for recording the physical features of the snow cover, particularly for snowpacks in arctic regions. Snow cover magnitude, distribution and variability are important aspects and, according to the instructions, the defined layers are supposed to represent major storm-precipitation periods or periods of high wind drift. The stratigraphy is established by the observation of textural and hardness differences. Snow properties within the layers are measured in accordance to the International Snow Classification (Schaefer et al., 1954). In addition, a classification for snow surface conditions is established (CRREL, 1962). These guidelines are prepared for snowpack surveys on the Greenland and polar ice sheets because the investigation of the stratigraphy of polar snow and firn plays an important role in climate research. Layer boundaries are used as indicators of secular climate change. Benson (1962) traverses Greenland and uses snow and firn stratigraphy to establish the state of the mass balance of the ice sheet. The spatial variation of the snowpack aids in his reconstruction of storm events. For large scale observations (20-40 km) Benson (1962) develops a special glacial climate snowpack classification system, which is based on altitude and climate on the ice sheet.

Summary of period 1940-1960s

This period brings about the first International snow classification. The focus in this classification is on the properties of disaggregated snow, the grain size and grain shape. International grain shape classifications are brought together in a field guide to snow crystals. Quantitative textural analysis is further advanced in laboratory experiments. Snow surface characterisation is advanced and a snow surface classification is presented. New instruments to capture snow hardness and snow strength are developed and compared and the ram and hand test are combined. Translucent snow profiles illustrate the high variability of snow stratigraphy and snow properties in natural snow profiles and show that stratigraphic description is incomplete. A systematic analysis of snow properties from the translucent profile images was not advanced.

2.2.4 From 1970 to the 1980s

Descriptive methods

Systematic investigations of the mountain forest snow cover are carried out and documented in 1978 (In der Gand). Snow profiles are taken in lines in flat study plots with classical methods. In cross section drawings In der Gand illustrates the large vertical and horizontal variability of the mountain forest snow stratigraphy. In der Gand (1978) qualitatively relates the increased stability of the snowpack on forested slopes to the textural inhomogeneity inherent in these profiles. Mountain forest snow cover variability is further described by Imbeck (1983; 1987).

Optical methods

Dye method:

A simple optical snow stratigraphic method is developed by Kovacs (1976), who sprays the pit wall with dye. The method is much simpler than the bonfire method (Nakaya et al., 1936) but it does not produce as good contrast for photography.

Drawings, photographs:

Cross section drawings are used by In der Gand (1978) to document snow profiles of mountain forest snow covers. He illustrates the large vertical and horizontal variability of the mountain forest snow stratigraphy. Imbeck (1987) also records mountain forest snow covers using photographs and drawings of snow profile cross-sections (Fig. 16).

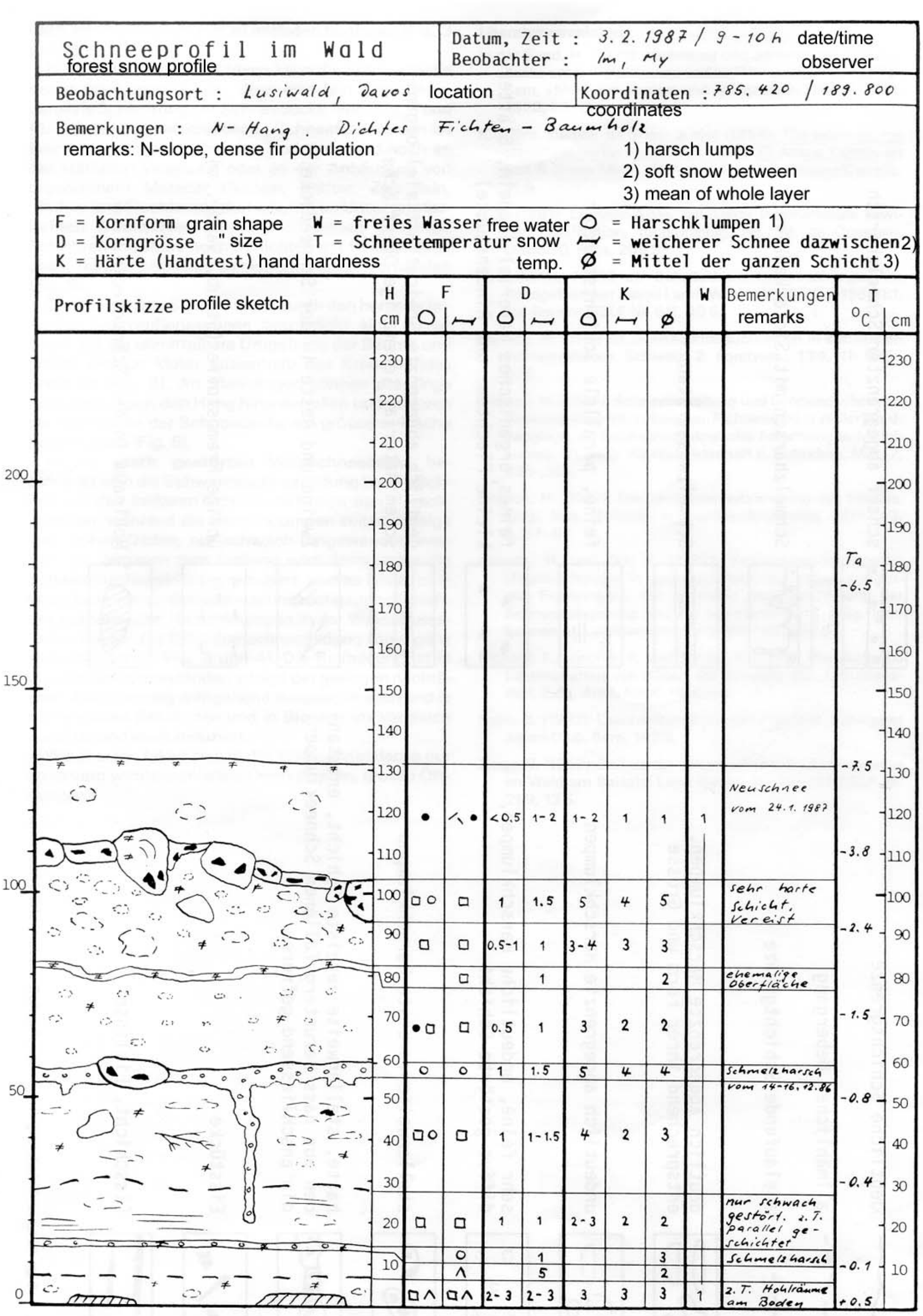


Fig. 16: Forest snow profile with a 2-dimensional profile sketch. A method developed and used by Imbeck (1987). The great variability in the profile and the not parallel layering required new morphological classes and more differentiations in layer boundaries.

Thin section, scanning photometer, low scan video, serial section analysis:
Good (1970) develops an automatic laboratory method to photograph and analyse snow structure from thin sections via the optical properties of the samples and presented numerical parameters to identify snow structure (Good, 1974). Furthermore, Good (1980) compares the methods of scanning photometer and low scan video technique for the analysis of the two-

dimensional snow micro structure. It turns out that only for half of the scanned snow structures the methods produces consistent results. Good (1982) uses thin sections to study snow stratigraphy of snowpacks on ice to help remove ambiguities in dating the snow layers. It is shown that the stratigraphy is greatly smoothed out in the firn and it is possible to verify the depth at which the transition from firn to ice occurred. With micro structural analysis, Good observes for the first time large density variations in artificial snow samples. Good (1987) continues his research on image analysis of snow thin sections, serial cuts and on the three-dimensional analysis of snow structure. From many micro structural parameters he finds the distance between grains the best single parameter to characterise the structure of different snow types, for which he shows a link to tensile strength. However, for irregular grain shapes, a three-dimensional analysis of the geometry of the snow structure is necessary (Good, 1989). In their study of the relationship of air-permeability (in terms of pore space or grain size) and strength, Buser and Good (1987) cannot find a meaningful relationship.

Thick section:
Harrison (1982) develops a thick section sampler for large snow profile samples. With this technique, bulk profile sections up to 1 m height are prepared for photography on a light table. Measurements of the snow properties can be taken immediately from the bulk sample. A comparison to the bonfire and dye-techniques shows (Harrison, 1982) improvements in the layer resolution of the snow profiles.

Radar:
The relationship between the electromagnetic scattering properties and the physical properties of the snowpack is first presented by Ellerbruch, Little et al. (1977) and Boyne and Ellerbruch (1979; 1980). With a ground-based FMCW active microwave radar system (8 to 12 GHz) they observe stratigraphic changes throughout the winter of 1978 and show that it is possible to measure snowpack water equivalence to ±5 % accuracy with this system. Gubler and Hiller (1984) develop a ground-based microwave FMCW radar system for snow and avalanche research. Density discontinuities are interpreted from the backscatter, but no quantitative correlation of amplitude to stratigraphy was done. Continuously changing properties cause multiple refractions that are difficult to interpret. In snowpacks with homogeneous density, such as polar snowpacks or alpine snowpacks in the melting phase, water equivalent estimations are possible (Fujino and Wakahama, 1985; Gubler and Hiller, 1984). The results demonstrate the potential use of active microwave systems for snowpack monitoring, but improvements in four areas are necessary: the relation between physical snow parameters and microwave frequency, the development of theoretical models as a function of snow parameters, improvements in the measurements and elimination of ambiguities from the analysis, adding a hardness profile to the analysis (Fujino and Wakahama, 1985).

Morphological methods
UNESCO (1970) issues guidelines on seasonal snow cover observations as part of the International Hydrological Decade which was later published in handbook form (UNESCO, 1981). These guidelines assist in an international inventory of snow and ice masses. The method of stratigraphic snow pit records is based on the International Snow Classification (Schaefer et al., 1954).
Armstrong (1977) studies snowpack characteristics and its relationship to avalanche release in the San Juan Mountains, Colorado. He shows that temperature gradient metamorphism was mainly responsible for the highly differentiated stratigraphy with low mechanical strength on

all their study sites. He is first to report on sub-surface metamorphism driven by solar radiation, which typically produced persistent weak layers in the snowpack. These layers were mainly responsible for failures when new snow loads accumulated. Armstrong (1977) resumes that to understand and predict avalanche occurrences, it is rather important to know the strength profile of the old snow structure next to knowledge about the new load. Ferguson (1984; 1985) also investigates the relationship of snow stratigraphy to avalanche formation. Her snow profile records are based on the guidelines of the International Snow Classification (UNESCO, 1981) but wherever possible her records are extended from discrete ordered data to continuous metric measurements for better quantification of the snow stratigraphy. The temperature profile is adjusted to mid-layer depths and layer boundary temperatures are also taken. From these measurements the temperature gradients for each stratigraphic layer can be calculated. Smaller density cutters help to get a higher resolved density profile. Measurements for thinner layers are omitted or integrated with neighbouring layers. Ferguson (1984) also defines a stress-scale to quantify hand hardness, but still cautioned about the subjectiveness and inaccuracy of the test. It is not well suited for layer hardness comparisons across snow pits (Ferguson, 1984). The ram hardness test is entirely omitted because of its inaptness to capture the essential thin, weak layers. Ferguson's (1984) extends grain shape classification by including information on the general metamorphic state, the degree of riming, the degree of rounding (destructive metamorphism), of building (constructive metamorphism) and of bonding (wet metamorphism). It results that all structural parameters are necessary to assess the avalanche potential of a slope. Ferguson (1984) concludes that quantitative snow profiles are a successful method, but improved measurements of snowpack properties are necessary to improve the assessment of snowpack stability.

Mechanical methods

Hardness measurements:

Snow mechanical advancements are achieved by St. Lawrence and Bradley (1973), who compare the resistograph to the ramsonde. They find a good correlation between these instruments and show a much better resolution of soft and thin layers than possible with the ramsonde. The resistograph is still not widely used, perhaps of the complicated mechanical procedure and its limited maximal force range. Gubler (1975) improves the ram hardness equation, yet he finds that the ram hardness cannot be correlated to the strength properties of the snow because its resolution is too low to account for the inter-granular snow structure responsible for the tensile strength. De Quervain and Meister (1987) review 50 years (1936/37-1985/86) of snow profiles at Weissfluhjoch and the relationship to avalanche activity. Conclusions about dominant factors responsible for the formation of weak and strong layers are drawn from the analysis. 50 ram profiles are classified. De Quervain and Meister (1987) present the first stability classification system from ram hardness profiles. It consists of six ram profiles types with specific stability attributes (Fig. 17). The analysis shows that snow stratigraphy is relevant because 25 % of all avalanche periods were attributed indirect reasons, which are in the snow stratigraphy. This result is confirmed by a further study of snow profiles taken near the rupture lines of avalanches (De Quervain and Meister, 1987).

Because reliable field measurements still lack the spatial resolution to capture potential failure planes, Föhn (1987a) uses the rutschblock stability test. It is a field test, which was

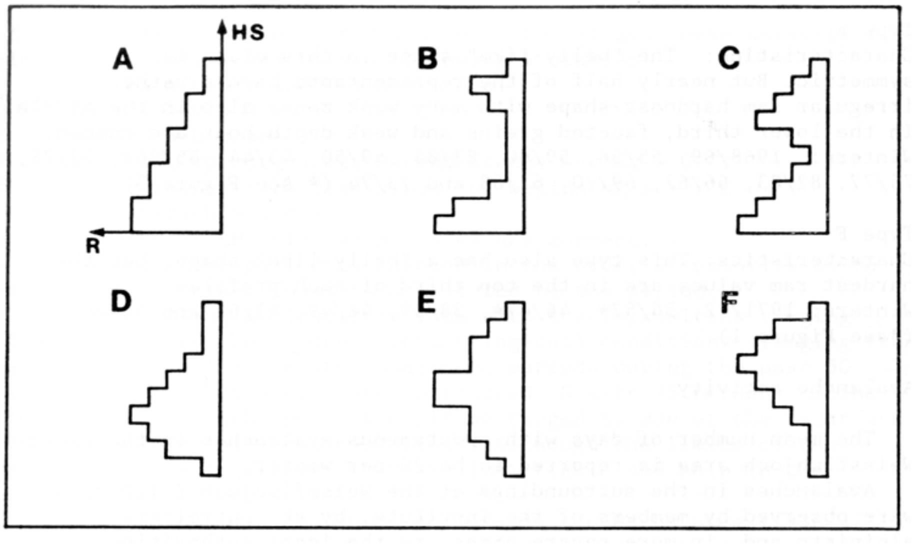


Fig. 17: Stability classification based on ram profiles with stability attributes (DeQuervain and Meister, 1987).

developed by the Swiss Army in the 1970s. The rutschblock test returns an index of the stability of the weakest area in the isolated snowpack column (1.5 x 2 m). Föhn correlates the rutschblock index with stability index gained from the shear frame test and with empirical avalanche activity observations. Föhn (1987b) describes special shear frame procedures and analytical methods to prove the validity of the stability index approach.
The digital resistograph is based on the idea of the resistograph (Bradley, 1966) and is re-introduced by Dowd and Brown (1986). This instrument records the force every 5 mm at variable speed, it allowed a faster, repeatable measurement with better spatial resolution than the ramsonde. However, layers thinner than 5 mm, often related to failures in the snowpack, cannot be captured by the digital resistograph. Frequent malfunctions and the lack of durability in the field are probably the reasons why it never became accepted. Brown and Birkeland (1990) describe a more developed prototype of the digital resistograph, with higher resolution and digital data storage and download functions. Though the results are promising, the durability of the force sensor and the electronics are problematic. The digital resistograph was not further developed since then. Schaap and Föhn (1987) test a penetrometer developed from a commercial geotechnical instrument, with a cone diameter of 11.3 mm and a 60° included angle. It has a spatial resolution of 1 mm in hard snow and the data is recorded on a chart recorder. They compared the cone penetrometer hardness profile to the ram hardness profiles (Fig. 18) and the new penetrometer shows greater variations in the snow stratigraphy and snow hardness than the classical methods. No other reference profiles are available and therefore, the signal interpretation is difficult. Despite promising results the cone penetrometer was not further used in snow research. Other approaches are followed by Abe (1991), who introduces a fibre optic snow layer sonde with multiple sensors for snow stratigraphic measurements but no further measurements or verifications have been shown so far. The PANDA (*P*enetrometre *A*utonome *N*umerique *D*ynamique *A*ssiste par ordinateur) is developed for soil mechanical studies (Navarre et al., 1994). It is a dynamical cone penetrometer with electronic registration. As with the ramsonde, the resistance to penetration of snow can be determined by observing the amount of penetration after each (hand held) hammer drop. The energy input and the penetration resistance are registered. Energy losses

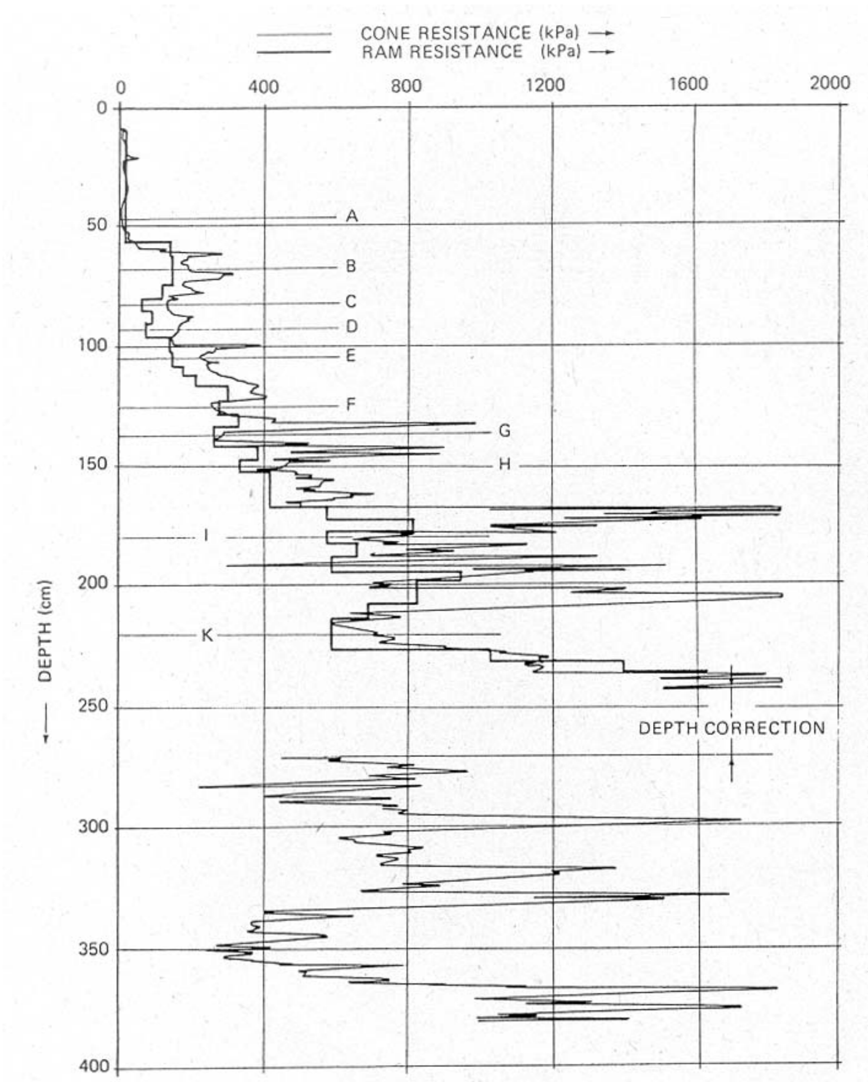


Fig. 18: Cone penetrometer hardness profile compared to ram hardness (Schaap und Föhn, 1987). The cone penetrometer resolves more layers and greater hardness variation in the snow profile than the ram test.

are not accounted for, and the spatial resolution is about 10 mm in sufficiently hard snow. Similar to the ramsonde, the spatial resolution is much lower in soft snow and thin layers.

Systematic investigations of the spatial variability of the snowpack and in particular its relation to snow cover stability are carried out by Conway and Abrahamson (1988) and Föhn (1988). Conway and Abrahamson (1988) find many small deficit areas (layers of small spatial extent with little strength) in the snowpack on slopes which are related to avalanche formation. Föhn (1988), however, finds continuous deficit areas of large spatial extent, and only these continuous weak layers to be responsible for avalanche release. These two hypotheses start a new discussion about snowpack properties, snow stratigraphy and its spatial variability. Analytical snow fracture-mechanical models also presuppose a strength deficit area, otherwise initial failure and failure propagation cannot be explained in the existing model (Bader et al., 1989).

Textural methods

The relation of snow textural parameters with snow mechanical parameters regains attention by Keeler (1969), St. Lawrence (1974), Kry (1975a; 1975b) and Gubler (1978) who recognise that snow density only partially explains mechanical strength, because very different snow micro structures can exist at equal snow densities.

Imbeck (1983; 1987) records forest snow covers on a flat field and on a slopes and compares the profiles to snow covers in open areas. Snow interception by trees produces the great textural variability and the non-parallel stratification of the snow cover in forests. For this reason Imbeck develops a specialised snow classification (Fig. 19). Snow layer boundaries are distinguished between clear layer boundaries, continuous layer boundaries and boundaries, which gradually taper-off. Imbeck (1987) also includes classes for clearly and unclear defined lumps of harsch, lumps of ice and other disturbances in the profile.

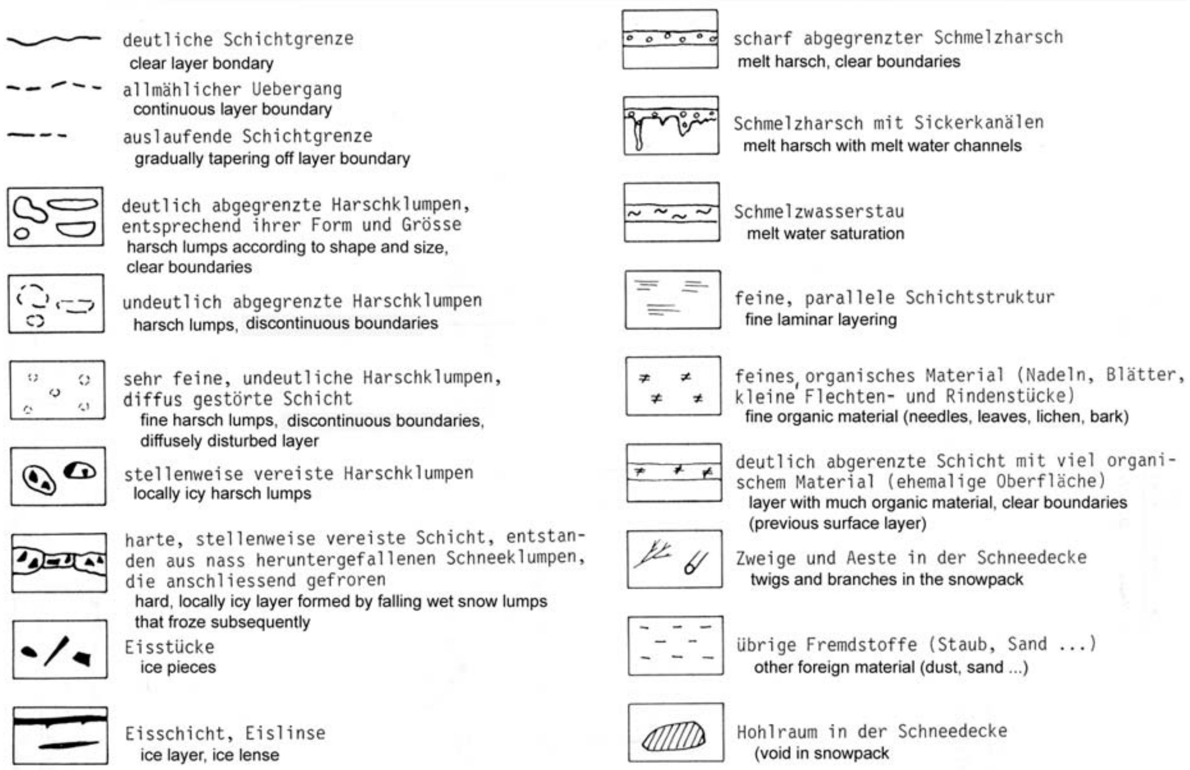


Fig. 19: Snow classification for mountain forest snowpacks (Imbeck, 1987). A new concept are the classes of layer boundaries.

Summary of period 1970-1980s

The undisturbed snow texture, which consists of the grains, their interconnections and their spatial arrangement gains larger scientific interest during the 1970s and 1980s. In order to capture and model the properties of the snowpack as a whole the analysis of the individual elements is insufficient. Snow mechanics and snow metamorphism need to be investigated in the snow textural context in order to understand the processes that lead to changes in the snow micro structure and cause instabilities in a slope snowpack. New methods are needed and become adopted from other disciplines. Seismic, optic and electro-magnetic methods, which are successfully used in geology are adopted in snow science. The availability of faster computers is particularly useful in the advancement of image analytical methods. Good

(1972) states that the measurement and parameterisation of the snow structure is a prerequisite to model the mechanical properties of snow and to eventually interpret the critical behaviour of a snowpack (instability, failure) from a snowpack model. In his history of snow cover research, Colbeck (1987) summarises that *“the success of the current effort will be determined in part by our ability to characterize snow as an assembly of particles”*. Types of snow pack measurements are developed and improved during this time period. The importance of the spatial variability of snow stratigraphy and snow properties is reconsidered but most available methods are only suited for point measurements and the local information needs much interpretation to describe the continuous state of the snowpack.

2.2.5 From 1990 to present

Optical methods

Dye method:

In-situ records of the stratigraphy of the whole snow profile are improved by Kovacs (1993). He advances his profile wall dyeing method presented earlier (Kovacs, 1976). With the new profile preparation and dyeing method the resolution of the snow stratigraphy is enhanced but no quantitative interpretation or relation to snow properties is available.

Translucent profile:

Good and Krüsi (1992) refine the method of preparing translucent snow profiles in the field. With a heated wire the profile can be cut without disturbing or breaking weak layers or hard inclusions. The latter often occurs when the profile is brushed, scratched, carved or cut. Once this delicate profile is prepared and photographed, the pictures are binarised and analysed by image analytical tools such as pattern recognition or linear structure analysis. Good and Krüsi (1992) judge these profiles better than dyed profiles, and the visibility of single layers as good as stereological images. While time consuming and delicate to prepare, translucent profiles are so far the only objective record of the complete and undisturbed stratigraphy of the snowpack (Fig. 20). Since the profiles are about 0.5 m wide, an insight into the small-scale spatial variability of the snowpack properties is also possible with this method. However, Good and Krüsi (1972) are unable to classify transmissivity to properties of the snowpack.

Surface section analysis, Micro CT:

The three-dimensional snow micro structure is measured and visualised by optical computer tomography from planar sections and x-ray micro tomography (Coleou et al., 2001; Schneebeli, 2000). Nuclear magnet resonance tomography (NMR) (Ozeki et al., 2000) is unable to deliver the spatial resolution necessary for natural snow. The methods are suitable to quantify the undisturbed snow micro structure. Disadvantages are the high cost and the small sample size. Schneebeli (2002) introduces a method of snow micro tomography, where besides the quantification of the three-dimensional micro structure of snow, continuous monitoring of the undisturbed sample is possible. He carried out snow metamorphic experiments with continuous measurements of the snow micro structure and heat conductivity (Schneebeli, 2002).

Radar:

Radar technology is used to investigate mountain forest snow covers by Gubler and Rychetnik (1991). Fig. 21 shows three FMCW radar profiles of snowpacks in an open field, in a mountain larch stand and in mountain spruce stand. The layering disappears in the radar

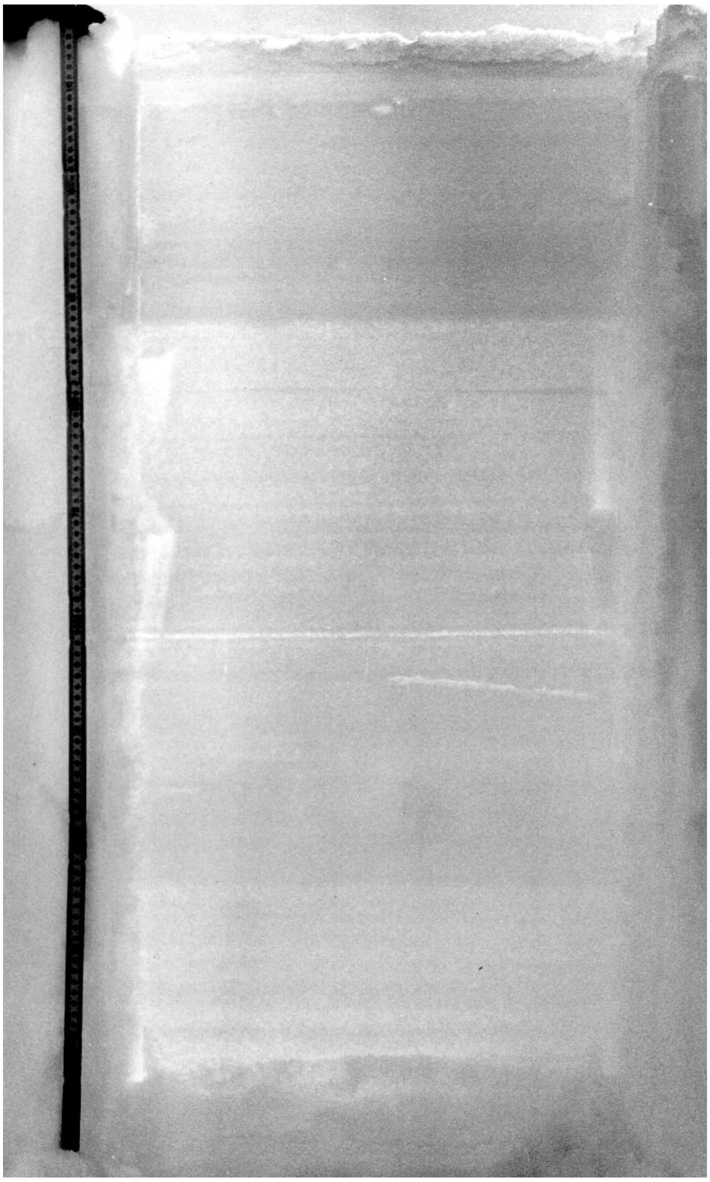


Fig. 20: Translucent profile with heated wire cutting, taken by Good and Krüsi (1992). The image analysis of these profiles produces as good results as stereological laboratory analysis and better results than the analysis of dyed profile images.

signals of the forest stands. Irregular snowpack disturbances, like harsch and ice lumps as described earlier by Imbeck (1987), do not appear in the radar profiles. Irregular shapes and properties produce multiple scattering of the electromagnetic waves within the snowpack, which makes such structures invisible in the resulting profile. In polar snowpacks, where the layering and the snow properties are very homogeneous, this method is applied more successfully (Foster et al., 1991) and it is possible to partially reconstruct stratigraphic properties. However, the vertical resolution is relatively low (dm–m). The inter-annual accumulation and layer thickness on glacial ice can be reconciled from the radar profiles. Further FMWC radar (26.5-40 GHz)) investigations in snowpack stratigraphy are carried out by Koh (1993) who can partially explain snowpack stratigraphy and spatial and temporal

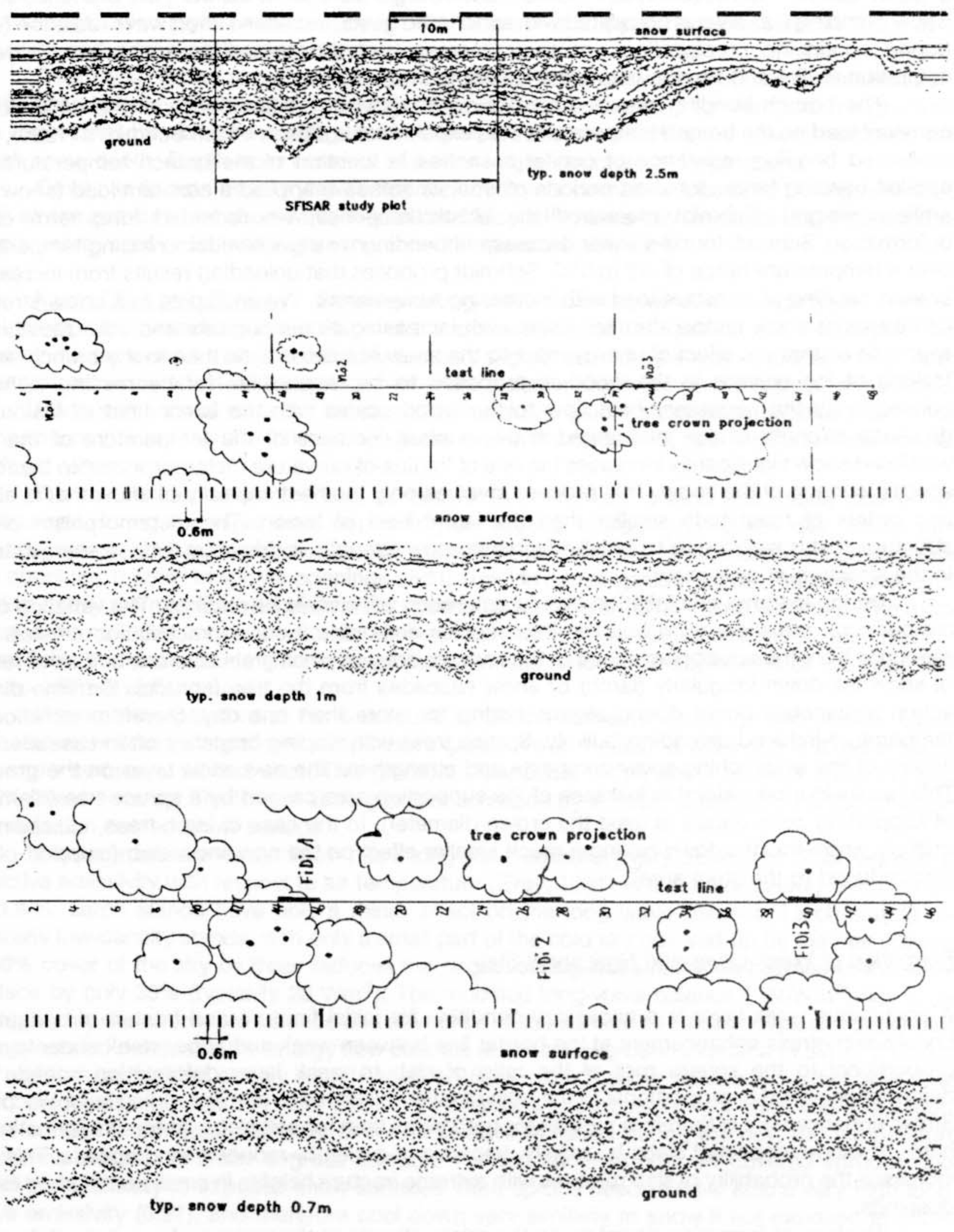


Fig. 21: FMCW radar profiles, from top to bottom: in an open field, of a loose larch stand, and of a dense spruce stand (Gubler and Rychetnik, 1991). The parallel layering disappears in the forest stands. Disturbances like harsch lumps steming from the trees act as scatterers and have a great influence on the backscattered signal.

variability from the radar signals. However, Koh cautions that a complete snow stratigraphy is not possible with the method so far. Koh (1993) suggests to use multi-frequency FMCW radars to determine its usefulness in snow stratigraphy. With improved FMCW radar technology, Holmgren, Sturm et al. (1998) approach the investigation of snowpack stratigraphy. Fig. 22 shows a shallow arctic snowpack in a two-dimensional radar signal cross-section. Only large density continuities produce a distinct radar backscatter, and if the resolution were increased, penetration depth and ground information are lost. Holmgren, Sturm et al. (1998) conclude from this study that snow stratigraphy cannot be resolved with the currently available radar technology.

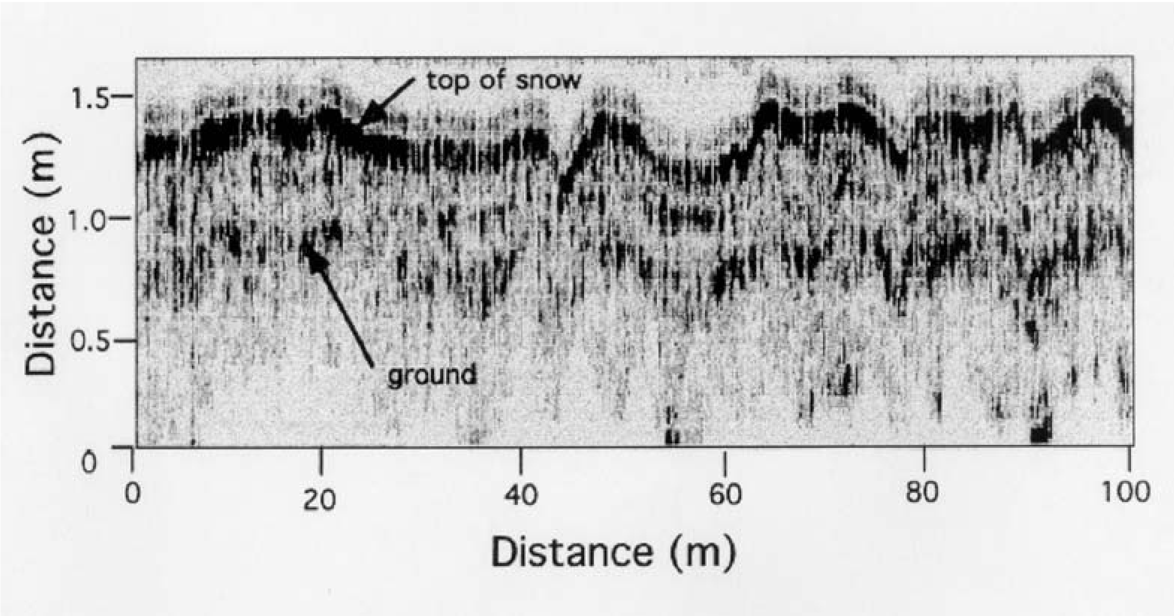


Fig. 22: FMCW radar profile of arctic snow pack (Holmgren, Sturm et al. 1998). It is not possible to resolve the complete snowpack stratigraphy. With higher resolution penetration depth and ground information are lost.

NIR photography:

Haddon, Schneebeli et al. (1997) introduce analog near-infrared photography (NIR) in snow profile research. Using quantitative image analytical procedures, layer boundaries are retrieved from the NIR images. Matzl and Schneebeli (2002) advance the method of profile preparation and introduce digital NIR snow profile photography. From the advanced images it is possible to retrieve the classical snow stratigraphy and a relationship of reflectivity (NIR-image grey values) and optical grain size is established.

Morphological methods

The new International Classification for Seasonal Snow on the Ground, prepared towards the end of the last period, is issued in 1990 (Colbeck et al.). It is a standard classification with focus on the physical snow properties in discrete, homogeneous snow layers. Measured parameters are mean grain size, grain size distribution, snow crystal morphology, bulk snow structure, and density. The classification of the snow properties is refined and new instruments and methods are taken into consideration. The classification of the snow texture is not included, Colbeck, Akitaya et al. (1990) state, *"Automated texture analysis could not be* *included in the International Classification due to the lack of a standard and unambiguous set of parameter definitions".* Colbeck (1991) reviewed the formation and effects of the layered snowpack describes layer formation processes and the effects layers have on the physical and chemical processes within the snowpack. Mean properties of the whole snow-pack or of the bulk structure are not sufficient. The smallest elements in snow stratigraphy, like thin crusts can be the controlling component for fluxes and forces and for the response of the snowpack (Colbeck, 1991).

A numerical energy and mass model chain for snowpack simulation (SAFRAN-CROCUS-MEPRA) as tool for avalanche forecasting is presented by Brun, Martin et al. (1989), Giraud (1992) and Brun, David et al. (1992). It calculates representative snow morphological properties for regions in the French Alps as a function of past and present weather conditions. According to Brun, David et al. (1992), the model is suited to efficiently and accurately simulate the evolution of the regional snow cover stratigraphy for a whole winter season. Fig. 23 is an example of a snow profile verification of a MEPRA simulation (Giraud, 1992). Lesaffre, Pougatch et al. (1998) determined objective grain shape characteristics from images of snow grains.

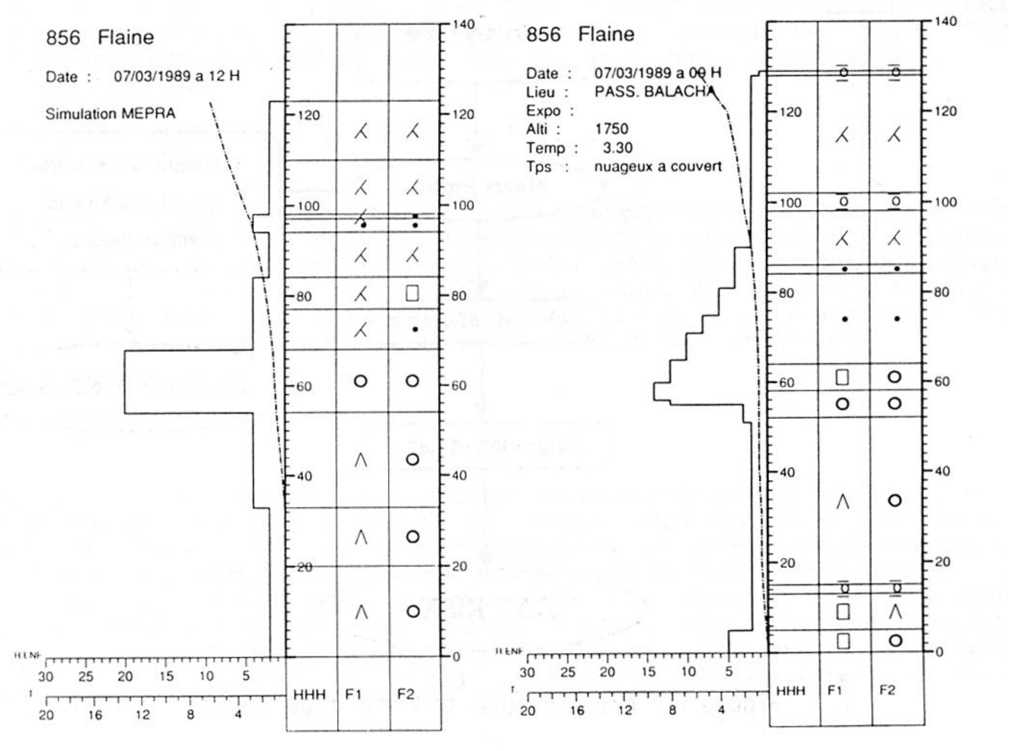


Fig. 23: Comparison of a MEPRA snow profile simulation (left) with a snow profile record (right) by Giraud (1992).

Fukuzawa and Akitaya (1993) investigate the formation of near surface faceting due to high temperature gradients. These thin layers with low mechanical strength form after snow deposition at the near snow surface because of the existing temperature gradients and their fluctuations. The influence of snow stratigraphy on snowmelt infiltration and wet snow metamorphism is observed by Albert and Hardy (1993) who compare differences between flat open sites and deciduous forest sites on slopes. Shultz and Albert (1998) present an automated procedure for plotting the morphologic snow stratigraphy, Fig. 24 shows the output of this graphical software. The illustration system is based on guidelines of the International Classification (Colbeck et al., 1990). It is unique because of the simultaneous display of many profiles which facilitates the qualitative analysis of temporal and spatial variability.

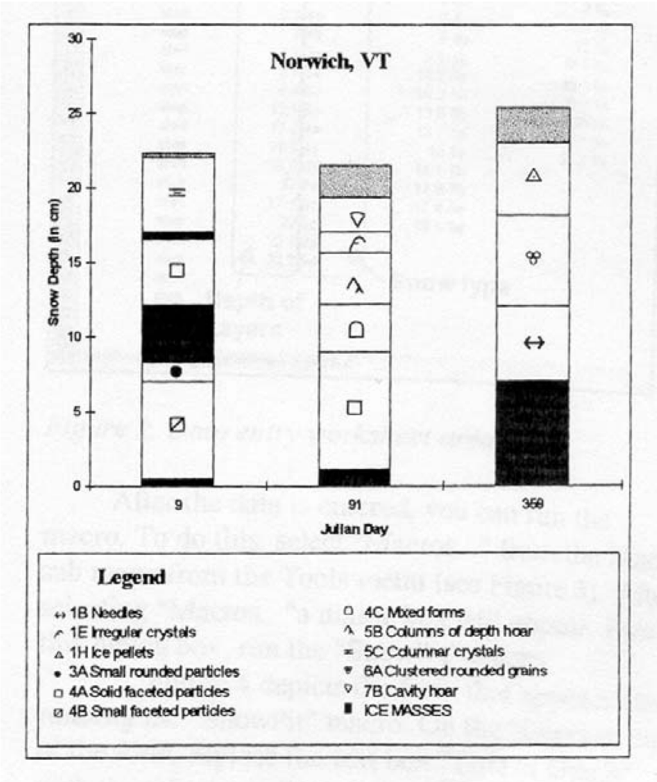


Fig. 24: Snow profile illustration from the SnowPit98 software by Shultz and Albert. The classification is based on discrete layers with homogeneous properties.

Mechanical methods

Hardness and strength measurements:

Föhn (1992) carries out a systematic investigation of weak layers and weak interfaces. The stratigraphic elements with low mechanical strength are a necessary condition for slab avalanche formation. To detect the weak layers and interfaces fresh avalanche fracture lines are inspected and stability tests (Rutschblock) are taken on potential avalanche slopes. Snow pit data and shear tests are taken from the snow profiles with failures. 40% of all detected weak zones are layers with a thickness of 1 to 60 mm, and 60% are weak interfaces where no distinct layer texture could be recorded with the classical methods (Föhn, 1992). This means that for snowpack stability investigations the traditional snowpack investigation methods are not sufficient in terms of the resolution of the mechanical and structural snow properties.

Birkeland, Hansen et al. (1995) investigate the spatial variability of snowpack stability. Depth and average snow resistance (an index of snow strength) are measured by the digital resistograph and related to terrain features on slopes. Snow resistance variations across a slab are recorded. The correlation to terrain features explains only partially the spatial snowpack variability. Snow over rocks is found to be significantly weaker than in adjacent areas.

Birkeland (1997) investigates snow stability and snow properties throughout a small mountain range in southwest Montana to better understand the spatial distribution of avalanches and improve their prediction. Stability is only weakly linked to terrain, snowpack and snow strength variables after relatively homogeneous weather conditions previous to data collection, and strongly linked after increasingly heterogeneous weather conditions. Generally, stability decreases on high elevation and northerly facing slopes.
To detect and measure fracture layers and their mechanical properties a variety of stability tests are introduced: The loaded column test is introduced by McClung and Schaerer (1993), the compression test (tap test) by CAA/NRCC (1995), the rammrutsch test by Schweizer, Schneebeli et al. (1995) and the stuffblock test by Birkeland and Johnson (1996). Birkeland and Johnson (1999) find a correlation of rutschblock and stuffblock results. Landry et al. (2001) introduce the quantified loaded column stability test.
Weak layer formation by the process of near-surface faceting and its relationship to avalanche formation is further investigated by Birkeland, Johnson et al. (1998). It turns out that weak layers of 1 mm thickness are formed within one to two days. Such layers remain as persistent weak layers in the snowpack and contribute to avalanche formation.
Stoffel, Meister et al. (1998) and Schweizer and Lütschg (2001) extend the stability classification, however this classification is still limited because it is based on the ram hardness profile and does not include thin layers. To gain an objective, highly resolved snow hardness profile, Schneebeli and Johnson (1998) develop a snow strength penetrometer, the SnowMicroPen, for field and laboratory measurements. The instrument's measurements are highly repeatable. The lack of subjective decision when operating the penetrometer makes the penetration resistance a quantitative measure of snow stratigraphy. The instrument dynamically measures the penetration resistance on a 5 mm diameter tip with a sampling distance of 4 μm. The high measuring frequency and small tip yield a measurement of single bond fractures (Schneebeli and Johnson, 1998). A comparison of the micro penetrometer hardness profile with the classical ram and hand hardness methods (Fig. 25) is presented for slope and flat field profiles (Pielmeier and Schneebeli, 2002). A comparison to surface section images of the same profiles shows that the classical profile records are incomplete and the snow properties in natural profiles are more variable than recorded in hand and ram hardness profiles. Kronholm et al. (2002) present a systematic investigation of the variation of snow stability and relate it to avalanche formation. MacKenzie and Payten (2002) introduced a new light, digital field snow penetrometer, the Sabre.

Textural methods

The geometric parameters gained from image analysis of translucent profiles are related to snow mechanical properties (Good and Krüsi, 1992). However, the link to a direct mechanical measurement is not made in this approach. Nohguchi, Ikarashi et al. (1993) consider the endlessly repeated finer structure in the stratified snowpack and considered it from the viewpoint of fractals. The snowpack is a result of the fractals in atmospheric turbulence and the subsequent time variations of snowfall events. In this model of the layered snowpack the number of layers is related to the probability with which a new snowfall event occurs. In their model, the thickness of the layers is related to the probability with which a certain daily new snow depth occurs (Nohguchi et al., 1993). Conway and Benedict (1994) show that the complex stratigraphy and the spatial variability of the snow properties results in a channelling of snowmelt. The water penetration into a layered snowpack is also delayed in comparison to an idealized, homogeneous snowpack. Sturm, Holmgren et al. (1995) develop a climatic snow classification with six classes for the distribution of Northern Hemisphere snowpacks and their properties. The properties are layer sequence, layer thickness, snow

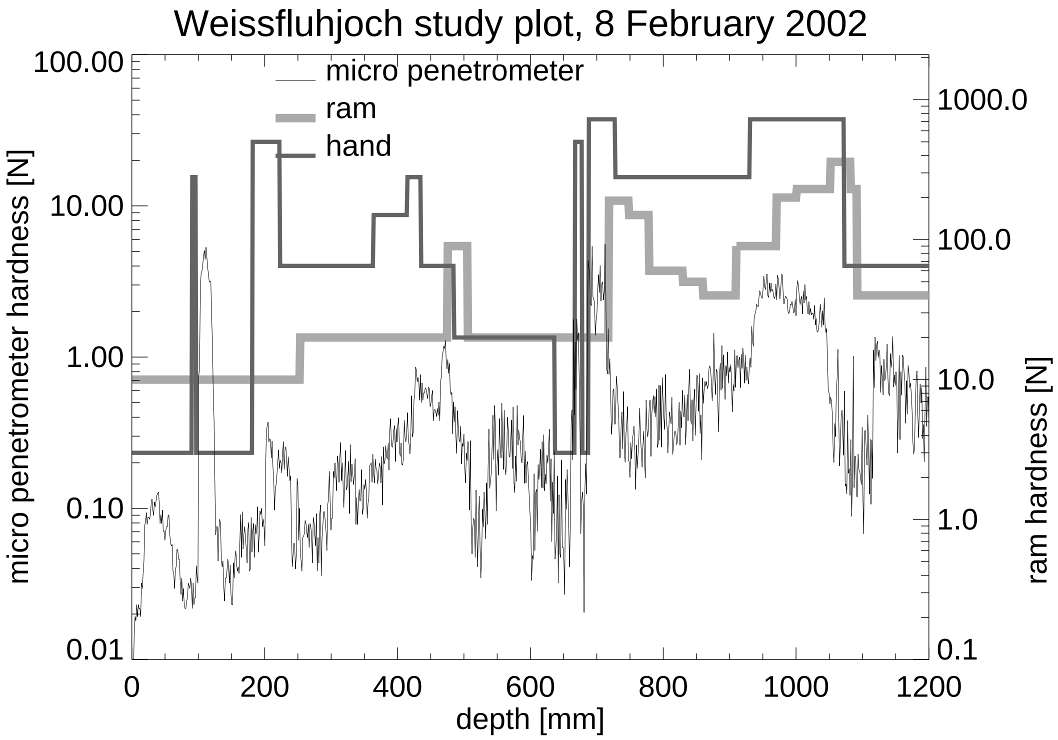


Fig. 25: Snow hardness measured by hand hadness and ramsonde (right y-axis) and by micro penetrometer (left y-axis, smoothed with moving average over 1 mm window). The logarithmic y-axes are chosen to get a better resolution in the lower hardness values. Differing in about one order of magnitude, the three hardness profiles can approximately compared at these scales (Pielmeier and Schneebeli, 2002).

density, crystal morphology and grain characteristics within each layer. The classification (Fig. 26) combines information on snowpack stratigraphy and on snow texture within clearly defined layers, within horizons without clear boundaries and within layers with variable properties. The classes can be derived from climate variables and classified snowpacks can be mapped for the use in regional and global climate modelling from climate data. Johnson and Schneebeli (1999) develop a theory of penetration, which is used to recover micro structural and micro mechanical parameters from the SnowMicroPen force measurements taken in different snow types. Laboratory experiments show that snow strength can be interpreted with a resolution of 1 mm and snow texture with a resolution of 4 mm from the micro penetrometer force signal (Schneebeli et al., 1999). A comparison of simulated (Lehning et al., 1999) and measured snow properties shows that the SnowMicroPen can facilitate the verification of snowpack models (Pielmeier et al., 2000).

Summary of period 1990-present

A new International snow classification is introduced, which is based on a morphological classification of snow in homogenous layers. Despite earlier quantifications and classifications of snow textural parameters, this is not included in the new classification. In search of further textural parameters, in-situ measurements of snow stratigraphy become increasingly important. Radar investigations turn out to be inapt to capture the snow stratigraphy. The method of translucent profiles and dyed profiles is advanced, so that the

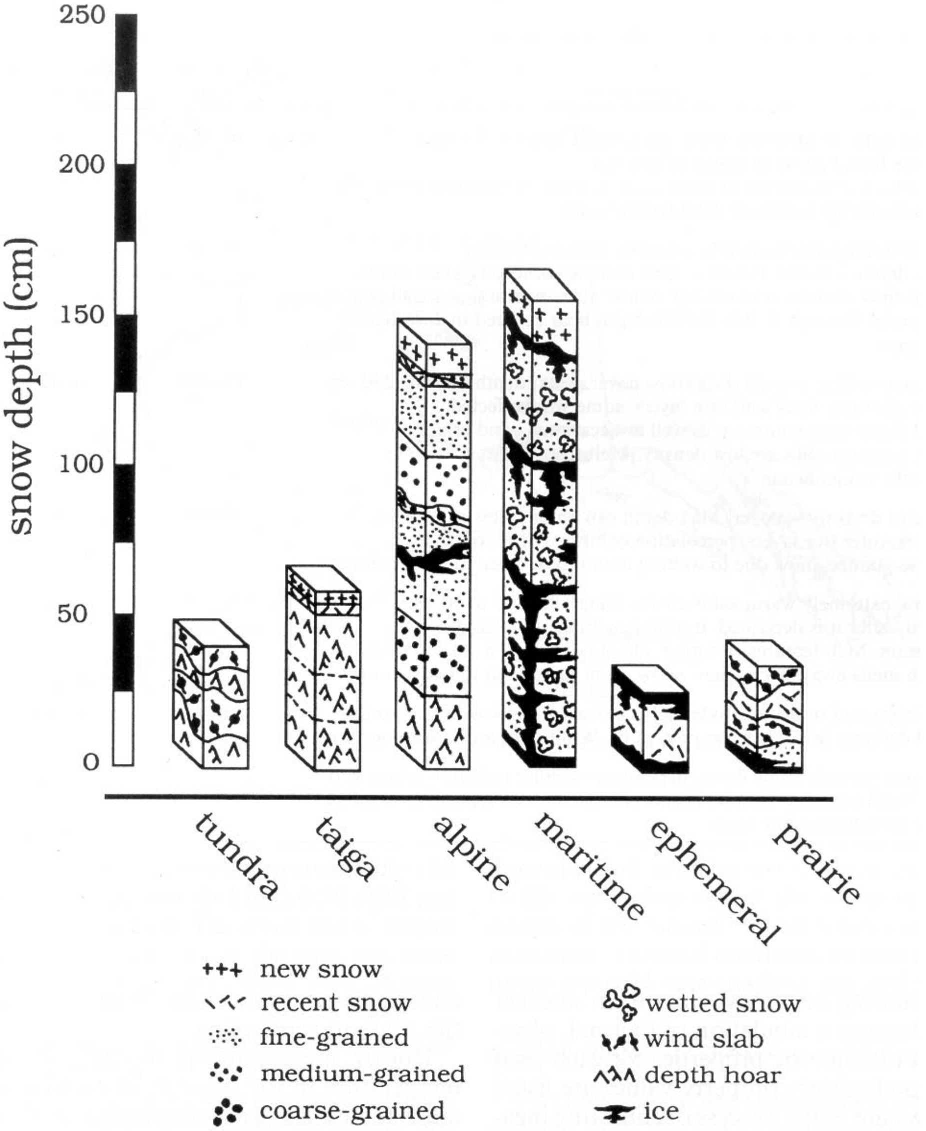


Fig. 26: Climatic snowpack classification system by Sturm, Holmgren et al. (1995) with classes for the seasonal snow cover according to stratigraphic and textureal attributes.

layering can be recorded with higher resolution. However, a quantitative analysis of these profile images is not available. New methods for snow micro structural analysis are introduced, they are x-ray and optical micro tomography. With these methods the three-dimensional snow micro texture can be quantified. The relation of micro textural to micro mechanical properties is now possible. For field applications, a variety of stability tests are introduced to gain stratigraphic and mechanical information about the weakest layer in a snowpack, information that is lacking in classical snow stratigraphy. The spatial variability of the snow stability is investigated in a few inconclusive studies that are carried out on snow stability. The importance of the high variation of snow stratigraphy and snowpack properties starts to be recognised. A new micro penetrometer for fast and objective snow hardness measurements is introduced and successfully applied in laboratory and field investigations. The highly resolved stratigraphy and hardness profile as well as the textural profile are available from the micro penetrometer measurements. The method of near-infrared photography of snow profiles is introduced. Continuous snow stratigraphy can be quantified from the images and the reflectivity is related to grain size.

2.3 Requirements for a modern snow stratigraphy

The stratigraphy of snow must cover different spatial and temporal scales. How far is this goal realized by current methods? The spatial scale at the surface can be covered by near-infrared photography and remote sensing, but cost may be quite high (Painter et al., 2001). Typical digital cameras have now a resolution of more than 2000^2 pixels, the grain and eventually the grain type distribution of a snow field of 100 x 100 m can be resolved with a spatial resolution of 0.05 m, a small scale plot (10 x 10 m) down to a 10's of grain clusters. Vertical snow profiles can be done with a speed of 0.5 m / min, with a spatial resolution of 0.5 mm and texture recognition. Undisturbed surface imaging by radar will be a challenge by using new very high frequency radars (30 GHz), where a theoretical resolution of 0.005 m should be possible, at least in dry low density snow. However, radar will never measure properties like strength of bonds, and may develop in the near future to be a help to track layers, but not their texture. The strategy should focus on sensor integration, combining several high resolution methods digitally. This will also require the use of new positioning techniques such that the properties can be geo-referenced. Numerical modelling of snowpacks should abandon the old manual classification methods as reference, and move to more objective and more physical methods. According to the Defense Modeling and Simulation Office (DMSO, https://www.dmso.mil/public/resources/glossary) of the US Department of Defense, validation of these models requires an *"accurate representation of the real-world from the perspective of the intended uses of the model or simulation"*. Most of the current methods used are only partially fulfilling this requirement.

2.4 Summary

Snow stratigraphy is a highly complex spatio-temporal system and the most promising efforts today are the instrumental and analytical developments that yield relevant physical, quantitative and verifiable snow parameters such as texture and strength. They also take the high vertical, lateral and temporal variability of the mountain snowpack into account. A combination of the physical approach developed in the last century with the geomorphologic-sedimentologic approach from the very beginnings of snow stratigraphic research could greatly enhance the understanding and interpretation of snow stratigraphy and snowpack properties. With the highly advanced technical methods available since the last decade, a combination of these two approaches is now possible. Highly resolved textural and mechanical measurements provide the accurate representation of the snow stratigraphy and snow properties. These sources can greatly improve process studies and process simulations of the snowpack. Additional spatial variability investigations will show whether scaling laws can be established, and whether the snowpack can be modelled spatially. Sensor integration will be a necessary next step to view the snow stratigraphy. The recent developments in snow stratigraphy challenge the general assumption of a snowpack consisting of discrete layers with homogeneous properties. The classical texture approach could be replaced by a properties approach, giving more insight into the mechanical, thermal and hydrologic behaviour of snow.

2.5 Acknowledgement

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Chapter 3

Measuring snow profiles with high resolution: Interpretation of the force-distance signal from a snow micro penetrometer

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Abstract

The classification of snow profiles is traditionally based on the discrete interpretation of snow layers in a snowpack. The layers are discriminated according to the physical properties of the snow. To account for the micro mechanical behaviour of snow in deformation, the classification needs to be extended to the micro structure of snow. A new instrument, the high-resolution snow penetrometer "SnowMicroPen" has been introduced. It is a new instrument to continuously measure the snow hardness at the micro scale. The new method of snow interpretation is described. Different quantitative approaches to interpret the penetrometer force-distance signal are shown. The penetrometer force signal reveals the great variability of the micro and meso properties in a snowpack. A comparison of the physical characterization of snow and the new, mechanical characterisation is shown. Being able to measure the complexity of the micro-properties of snow is the basis for a snow classification that includes the deformational behaviour of snow. It is also the basis to investigate scale and spatial variability questions.

3.1 Introduction

The discrimination and classification of snow layers is the classical method to investigate the snow cover. La Chapelle (1992) reviewed various systems of classifications of snow. All the classifications described there are based on the description of types of snow crystals, how they are formed, and how they change after they have been deposited. The International Classification for Seasonal Snow on the Ground (Colbeck et al., 1990) is also a physical interpretation of discrete snow layers. Measures are mean grain size, grain size distribution, snow crystal morphology, bulk snow structure, and density. As Colbeck et al. (1990) state, "Automated texture analysis could not be included into the International Classification due to the lack of a standard and unambiguous set of parameter definitions”. A climatic classification system was developed by Sturm et al. (1995). This classification can be derived from climate variables. Snowpacks are also classified according to the textural and stratigraphic characteristics of the layers. These are the sequence of the layers, the thickness, density, crystal morphology and grain characteristics within each layer. The physical description of snow and snowpacks does not include information on the micro structure and the bonding of snow. Bader et al. (1954) recognized that the classical snow profile and traditional mechanical tests can provide some characterization of the mechanical properties, but often are not suited to record the data at the needed resolution, i.e. at the micro level. Subjectivity of the interpretation and lack of standard of parameters are further problems at hand. Shapiro et al. (1997) made this claim more recently. They stated that the measurement and interpretation of the snow micro structure is necessary to study the deformational processes in snow under loading. Knowledge of the behaviour of snow in deformation is fundamental to understand the processes of avalanche release and many snow engineering questions. Therefore, Shapiro et al. (1997) see the need to extend the physical characterization to micro structural properties that most influence deformational processes. Such a deformational classification requires a method to measure suitable properties of the micro structure of snow, such as stress, strain and strength. With this data, constitutive relationships of snow deformation could be established.

In soil research, Meigh (1987) and Huang et al. (1993) showed that penetration strength is correlated to soil mechanical properties. Fukue (1979) showed this correlation in snow. Schneebeli and Johnson (1998) concerted their efforts to develop a snow micro penetrometer, the SnowMicroPen that measures micro level snow grain bond ruptures. This new instrument is a constant-speed, small tip, high resolution snow penetrometer that samples the penetration force at a resolution of every 4 μm. It provides a unique force-distance signal for different snow types and Johnson and Schneebeli (1999) developed a statistical theory of penetration that recovers micro properties of these snow types from the signal. Schneebeli et al. (1999) showed that the signal of the penetrometer can be correlated to micro mechanical and micro textural properties of snow. They developed a semi-empirical model to predict the micro-texture of snow from the penetration force-distance signal. Pielmeier et al. (2000) suggested that penetration tests with the SnowMicroPen will greatly enhance the snowpack model verifications.

Several quantitative methods to interpret the force-distance signal have been developed. After we give a brief overview of the new instrument, the signal interpretation methods are illustrated. The definitions of the scales at which the one-dimensional snow profile is investigated are the:

- micro scale: snow bonds and grains
- meso-scale: snow layers
- macro-scale: snow profile

At all scales, geostatistical analysis is used to analyse spatial continuity. The texture index (Schneebeli et al., 1999) delivers a micro mechanical and micro textural classification at the micro and meso scale. The original force-distance signal is analysed at all scales. Spectral analysis of the force-distance signal is a tool for the quality control of the measurement. A direct comparison between the classic and the new method of snow profile interpretation is not possible. This is due to the quasi-continuous nature of the penetrometer measurement, where the sampling distance is significantly shorter than the size of the smallest unit in snow. A comparison of methods is made to show the additional information gained from the force-distance signal about snow micro structure and its variability. Finally, an outlook on the two- and three-dimensional measurement and analysis of the micro penetrometer measurements is given.

3.2 The SnowMicroPen

Starting in 1995 Johnson and Schneebeli worked independently to develop a penetrometer with high spatial resolution that could detect fine layering and micro structural effects. Their independent efforts have been combined to eventually develop the SnowMicroPen, seen in Fig. 1. The instrument can measure snow at very high resolution and it gives information on the mechanical properties at the micro scale. The fundamental concept is that a continuously recording, small diameter penetrometer will make a more direct connection to micro properties than do large diameter cone penetrometers (Johnson and Schneebeli, 1999).
The maximum length of a penetrometer measurement is 160 cm and the measuring speed is 20 mm s^{-1}. The penetration resistance is recorded dynamically at a sampling distance of 4 µm. The force sensor has a very high measuring range (-500 N to 500 N), which allows to cover the whole spectrum of hardnesses that may occur in a natural snowpack.

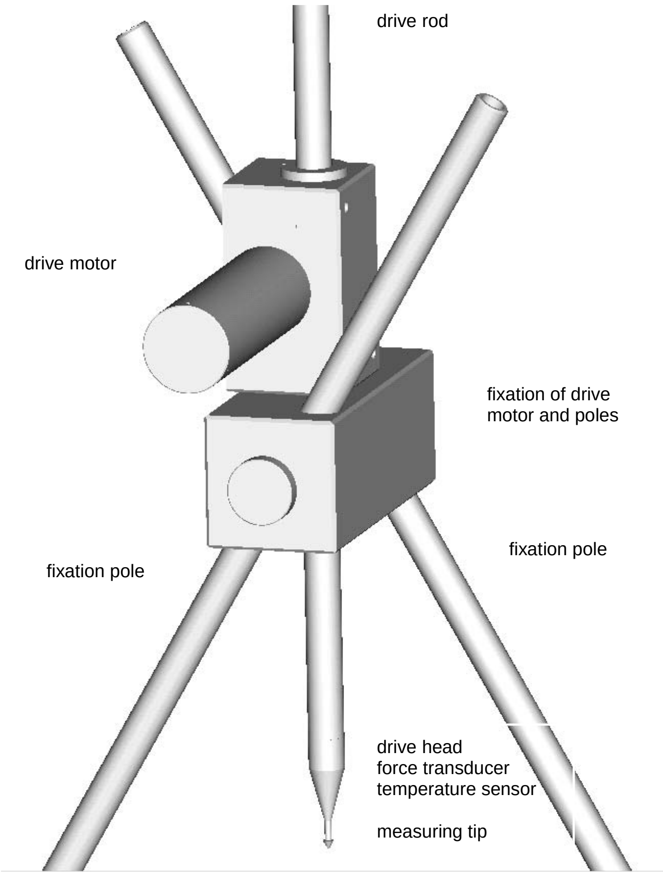

Fig. 1: View of the SnowMicroPen

During the last year, the SnowMicroPen has been advanced for field applications and the device was built more robust. Also, the operational software has been improved. A temperature correction is used to offset the drift of the force signal that occurs if the sensor experiences a temperature gradient during the measurement (Schneebeli, unpublished).

3.3 Interpretation of the penetrometer force-distance signal

In the following, the signal interpretation methods that have been developed are illustrated and the applicable scales are mentioned. The penetrometer force-distance signal is analysed here for one-dimensional snow profiles.

3.3.1 Analysis of micro scale texture using the original force signal

The micro penetrometer provides a unique signal for different snow types. The force signals in Fig. 2 and Fig. 3 are illustrated over a range of 4 mm. The measurement in Fig. 2 was taken in round-grained, equi-temperature metamorphosed snow. Statistical analysis of the signal gives a mean of 0.25 N. The distribution of the force values is symmetric about the mean and the signal is always above zero force. In contrast, Fig. 3 is an example of a force signal measured in faceted, temperature gradient metamorphosed snow. The mean force in this snow is at about 0.1 N, the distribution is positively skewed and zero force values reoccur throughout the measurement.

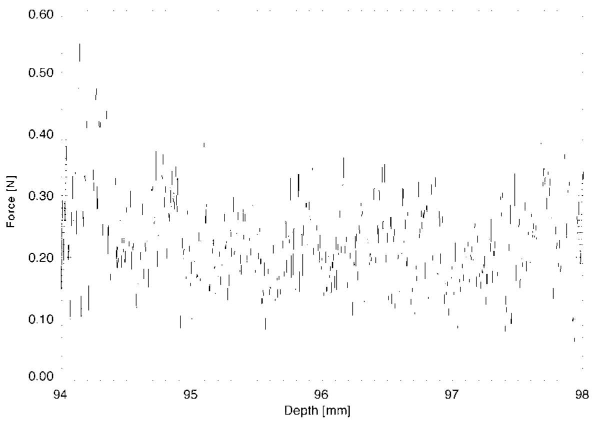

Fig. 2: Force signal measured in round-grained decomposed snow.

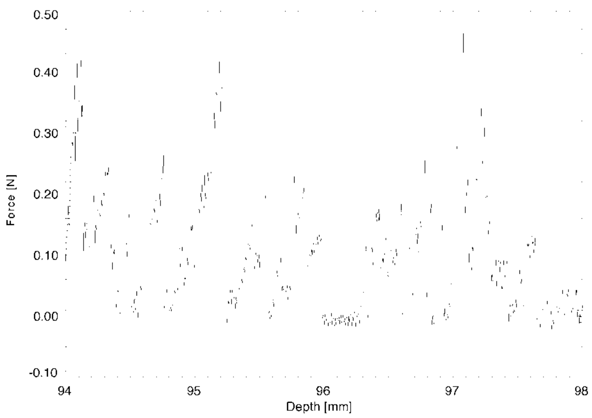

Fig. 3: Force signal measured in faceted temperature gradient metamorphosed snow.

The high resolution of the measurements provides information about the snow grain bond ruptures during the measurement. The different micro texture of the two snow types is clearly represented in their penetrometer force-distance records.

3.3.2 Analysis of the micro and meso scale spatial continuity using semivariance analysis

A tool to quantify spatial continuity in a measurement is the semivariance analysis. The force signal reveals the spatial continuity at the micro and meso scale. It is shown, how we can retrieve micro and meso textural information from this geostatistical signal analysis. Fig. 4 is the h-scatterplot for the round-grained snow sample in Fig. 2. The h-scatter plot is the force record at a distance $x+h$ (h=1 mm) as a function of the force record at a distance x (Isaaks and Srivastava, 1989). The distance between the spatially related force values is called the lag-distance or the lag. As the lag distance is increased, the similarity between pairs of values decreases and the points on the h-scatter plot spread out further from the diagonal line, where h=0.

The semivariogram summarises the relationships between all possible pairs of force separated by distances from 0 to a defined maximum distance, i.e. it is the summary of all possible h-scatter plots. As the paired force values on the h-scatterplot become less spatially related, the semivariance $\gamma(h)$ increases. The semivariance at a certain lag is calculated from the following:

$$\gamma[N^2] = \frac{1}{2n}\Sigma(x_i - y_i)^2 \qquad \text{for } i=1,....n$$

The semivariance at a defined lag distance is half of the squared difference between the force values and the 45-degree line, where $x=y$ on the h-scatter plot.

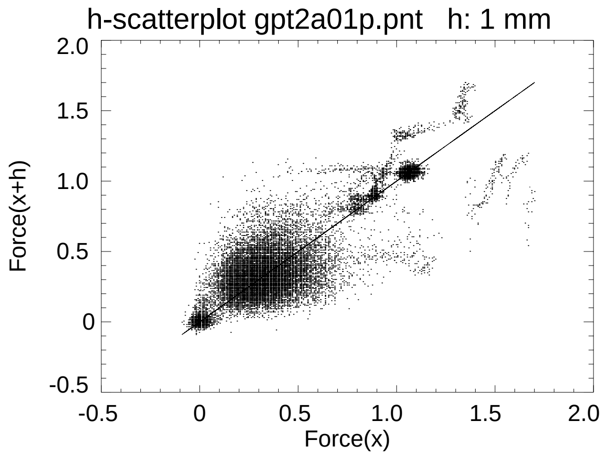

Fig. 4: H-scatterplot with a lag distance (*h*) of 1 mm. It shows the spatial relationship of the force values at the location *x* to the force value at the location separated by the distance *h*.

Eventually, an increase in the lag distance no longer causes a corresponding increase in the semivariance and the curve reaches a plateau, called the sill. Although the semivariogram for a lag=0 is theoretically 0, extremely short scale variability such as the noise of the force signal cause the vertical jump from the origin of the semivariogram to the value of it at extremely small distances. This is called the nugget effect. Fig. 5 shows the semivariogram for the unfiltered signal of a round-grained snow sample also shown in Fig. 2.

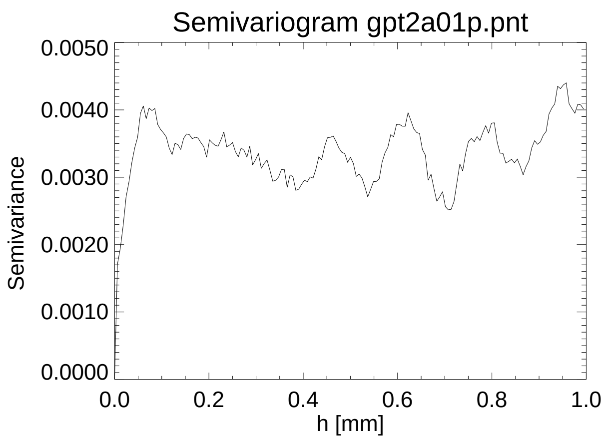

Fig. 5: Semivariogram showing spatial continuity up to a lag distance of 0.075 mm.

The semivariance is calculated with a lag distance of 1 point (which is equivalent to the sampling distance). The semivariogram is shown for a range of 1 mm. Spatial continuity at the microscale exists in this snow type up to a lag distance of 0.076 mm, which is equivalent to 19 force readings of the penetrometer. The noise of the signal is contained in this range. The sill of the variogram is reached at 0.0035 N^2. At a lag distance of 0.8 mm a minimum variance is reached. This corresponds to the interbond distance, which is assumed to correspond to the grain size in this well defined snow sample. When filtering the noise of the signal with a boxcar average filter with a length of 0.2 mm, the semivariogram reveals a larger scale of spatial continuity at about 4 mm, which can be attributed to some fine layering in the snow sample. Fig. 6 shows this second sill at about 0.001 N^2. The reduction of variance by a factor of about 10 by filtering the force signal shows that most of the variability stems from the noise.

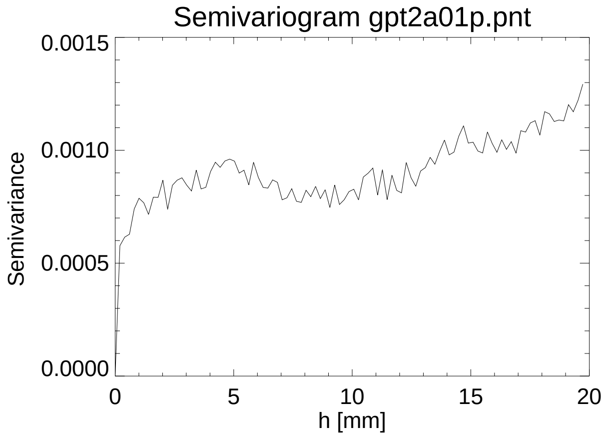

Fig. 6: Semivariogram with filtered force signal and reduced variance (for round-grained, equi-temperature metamorphosed snow).

The same analysis is performed on the faceted, temperature gradient metamorphosed snow as shown in Fig. 3. The semivariogram of the unfiltered signal is show in Fig. 7. The first sill is at about 0.75 mm and the first minimum is at 2 mm, which again corresponds to the grain size in this sample. The results from the semivariance analysis indicate micro-textural information. More samples need to be analysed and stereological analysis of bond properties need to be correlated to this signal analysis. Semivariance analysis of the force-distance signal is a new method to retrieve micro-textural information of the snow as well as about meso-scale layering in a snow sample.

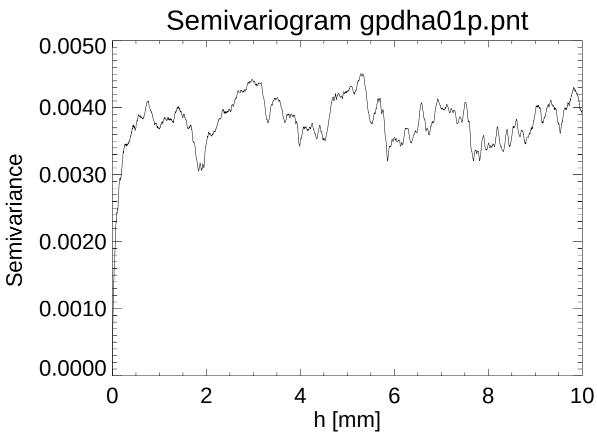

Fig. 7: Semivariogram with unfiltered force signal of faceted, temperature gradient metamorphosed snow.

3.3.3 Analysis of micro and meso scale snow hardness and texture using the texture index, and of discontinuities using the force signal

Based on the statistical signal analysis of the force signal, Schneebeli et al. (1999) developed a semi-empirical model to classify different snow types according to their micro mechanical and micro textural properties. Building on this textural classification of homogeneous snow samples, the interpretation is extended to natural, stratified snow profiles.

The texture index is the fraction of mean grain size (mm) and density (kg m^{-3}) and therefore a direct index of the volume density of grain contacts. It is correlated to the coefficient of variation of the SnowMicroPen force signal. The idea behind the texture index is that smaller spherical structural elements have a decreasing texture index, which is an indication of increasing textural stability.

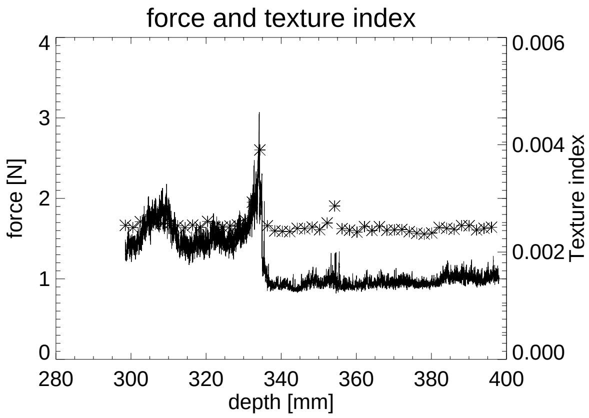

Fig. 8: Penetrometer force signal and texture index (asterisks) of a section of a natural snow profile .

Fig. 8 is the penetrometer force signal with the superposed texture index (asterisks) of a section of a natural snow profile. The profile section is of 100 mm and contains the interface of two bulk snow layers with a melt-freeze crust in the middle. The texture index (Schneebeli et al., 1999) classifies the snow hardness at a resolution of 1 mm and the snow texture at a resolution of 4 mm. Abrupt discontinuities in the force signal also produce a high texture index, as seen in Fig. 4. However, this cannot be interpreted in terms of snow texture. Such abrupt signal jumps must be classified as discontinuities in the snow texture. Therefore, the original force signal is evaluated simultaneously for abrupt force increases or decreases. The first derivative is calculated and if it exceeds a defined limit, it is classified as a discontinuity. The analysis also takes the snow hardness above and below the discontinuity into consideration.

3.3.4 Spectral analysis of force signal

The space-force signal is transformed into its space-frequency representation using Fast Fourier transformation (Oppenheim and Schafer, 1995). The Fourier transformation assumes an infinite, stationary signal. Since this is not the case, a small portion of the signal, i.e. a window function is used for the spectral analysis. The spectra of penetrometer measurements from 23 artificial and natural snow samples have been analysed. It results that the spectral signatures of the snow types are not significantly different and it is not possible to draw a classification from this analysis. However, the space-frequency representation of the signal gives insight into the quality of the measurement. Failures of the sample or resonance frequencies that occurred during the measurement may be hidden in the original force signal but can be easily detected in the space-frequency representation. Fig. 9 is the force signal of a snow sample and its space-frequency representation. Due to the great hardness of this type of snow the space-frequency plot shows resonant bands of high intensity, probably caused by a stick-slip interaction of the measuring tip or the cone with the snow. The resonant bands visible at the frequencies of 10, 20 and 30 mm^{-1} disappear at a depth of 65 mm where sample had an internal failure during the measurement. The resonant bands and the failure are only obvious in the space-frequency plot, which provides a quality control tool for the force measurements.

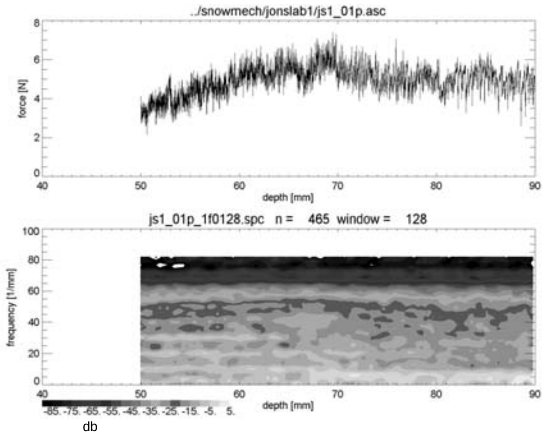

Fig. 9: Force signal and spectrum of a measurement where a failure occurred at depth 65 mm. The spectral bands seen above 65 mm disappear at this depth. Grey values show the intensities in decibel.

3.4 Comparison of methods: Classical snow profile, penetrometer profile, translucent profile

In this comparison it is shown, how stratigraphy is viewed in the classical snow profile as opposed to the penetrometer profile. The way a profile is measured and interpreted by the two methods is fundamentally different. To illustrate the layer discrimination process in the classical profile a translucent profile is shown also. The translucent snow profile is an isolated thin wall of snow photographed against the setting sun.

The classical snow profile is a one-dimensional record of the snow properties. A vertical profile wall is opened by digging a snow pit. The observer establishes the stratigraphy of the snowpack by sensing hardness differences by hand and inspecting textural differences by eye at a vertical line at the profile wall. For each layer the hardness and morphological snow properties are classified.

On 1 February 2000, the classical snow profile shown in Fig. 11 was taken at the Weissfluhjoch test field by the avalanche warning service, independently from the penetrometer and translucent profile. Previous to that day a two day snow fall period brought about 60 cm of new snow that was deposited on a melt-freeze crust. Below the crust, faceted crystals developed. The upper part of the profile is used for an exemplary comparison of penetrometer, classical and translucent snow profile. In particular we look at the section of the three layers marked in Fig. 10 and Fig. 12 with numbers, new snow (1), melt-freeze crust (2), second crust (3) and faceted crystals (4) . The box in the classical snow profile in Fig. 11 contains this profile section. Fig. 10 is the penetrometer force signal with the marked layers corresponding to Fig. 12. The penetrometer profile is shown for the section surrounding the melt-freeze crust. The depth 33 cm in the force profile corresponds to the snow height of 118 cm in the classical profile.

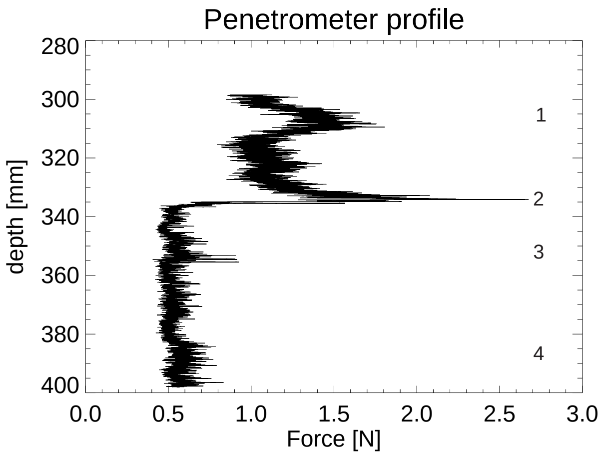

Fig. 10: Penetrometer signal of profile section. The positions of the numbers correspond to numbered layers in the translucent profile shown in Fig. 12.

The penetrometer force above the melt-freeze crust (1) has a trend to increasing hardness from 0.25 N to 1 N. The alternation of harder and softer layers in this part of the profile can be seen in great detail. The layering is due to grain size and/or density differences that could be a result of wind influence during deposition. This layering is less obvious in the translucent profile in Fig. 12 and hardly obvious in the hand hardness and ram hardness profile in Fig. 11. The melt-freeze crust (2) can be clearly interpreted from all profiles, however, the penetrometer measurement gives information about the exact thickness and the mechanical hardness of this very thin layer. The second crust (3), 20 mm below the first, is much weaker (0.9 N). In the classical profile it is not recorded. In the translucent profile the brightness of the layer resembles the one of the first melt-freeze crust, which might wrongly suggest similar properties. The snow below the second crust is faceted and has a larger grain size than the one above the crusts. This is reflected in the greater variability of the force signal (4). The comparison shows that the penetrometer profile agrees with the information from the snow profile and the translucent profile. Yet, the force signal gives much more precise information about the micro-properties of the snow and it reveals the great variability in mechanical hardness and texture within the layers.

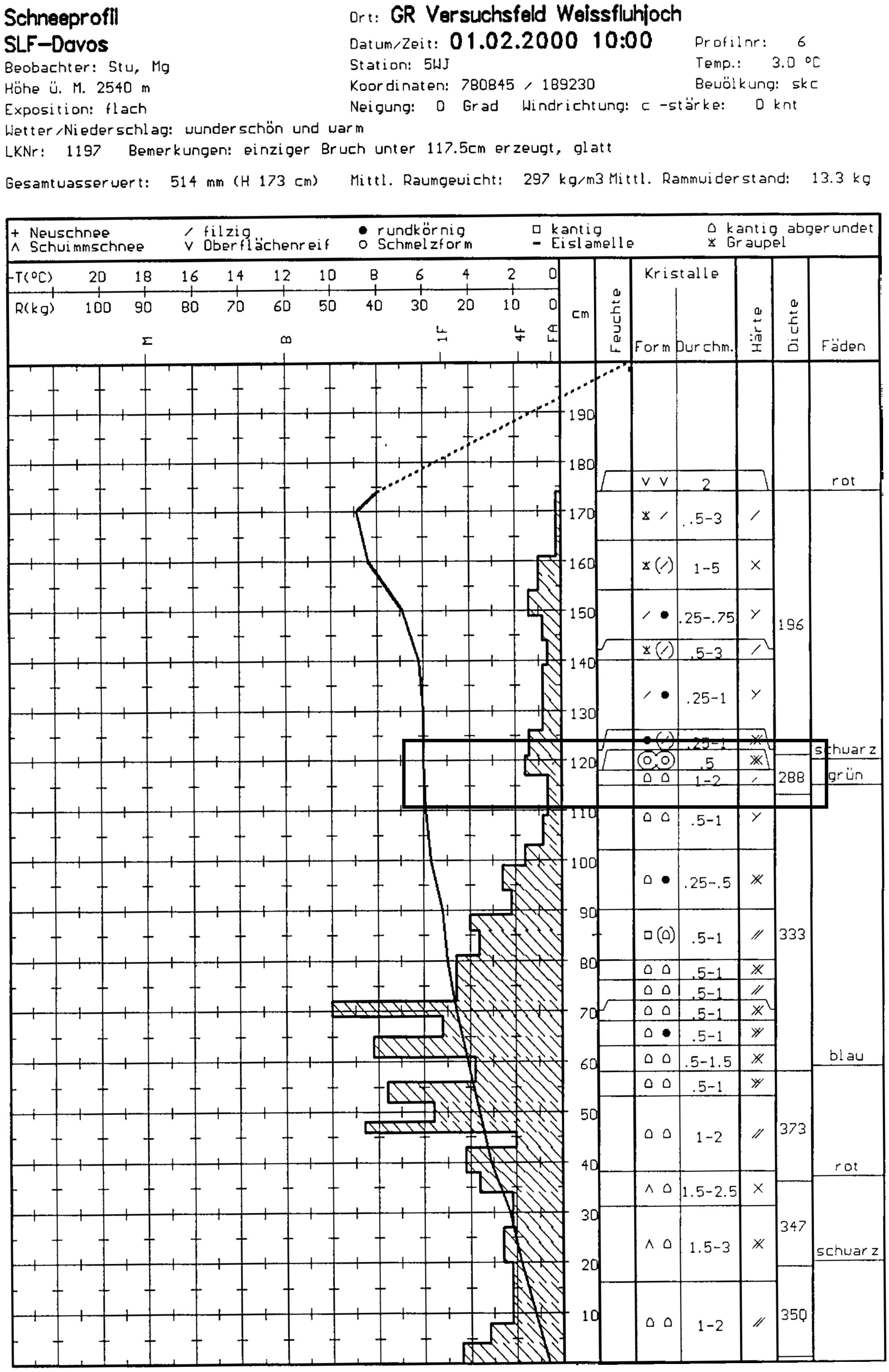

Fig. 11: Classical snow profile of test field profile. The profile section in comparison is contained in the black rectangle.

3.5 Results and discussion

The two different methods of snow profile interpretation are not directly comparable. Discrete layers that are defined in the classical profile are not always reproduced by the penetrometer. Moreover, the force signal shows the great variability in the mechanical and textural properties in a snowpack. Snow layers appear unequally clear in the penetrometer profile. The evaluation of the force-distance signal with the described methods provides a characterisation of the snowpack in terms of its micro mechanical and micro textural properties. The precise magnitude and mechanical hardness of thin layers and interfaces can be measured by the penetrometer. This new method of objective and quantitative snow profile interpretation is the first step to facilitate a quantitative analysis of the snow in deformation. More measurements need to be taken and a correlation to bond properties from stereological analysis will enhance the signal interpretation. The speed of the measurement allows to measure up to 120 snow profiles during a winter day. Hence, a systematic analysis of spatial variability at all scales is now possible. Systematic spatial measurements will be effected in the winter 2000/01. The analysis of spatial variability is combined with the analysis of snowpack stability and with the analysis of the micro properties of the snowpack. Now that we can measure snow micro structure it will be possible to manage the questions of micro mechanical behaviour of snow in deformation and the spatial variability of the snow micro properties.

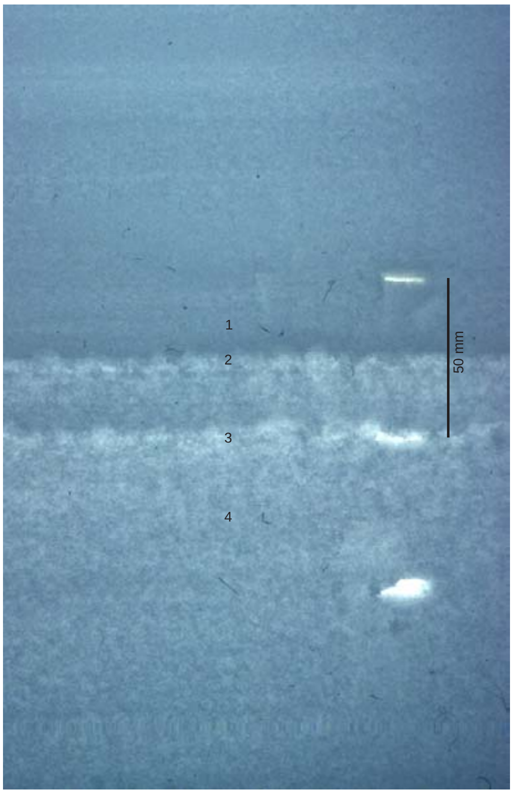

Fig. 12: Translucent profile of test field profile. The numbered layers correspond to the numbers on the penetrometer profile shown in Fig. 10.

3.6 Outlook: Visualization of two- and three-dimensional spatial variability

During the winter 1999/2000 the measurement of spatial variability of the seasonal snowpack with the SnowMicroPen has started. For the two-dimensional analysis of small scale spatial variability (cm to m) measurements have been taken in lines at the flat Weissfluhjoch test field. Fig. 13 is the visualisation of some of the two-dimensional force measurements. The vertical lines represent the actual location of the measurements. The spacing was increased from 100 mm to 200 mm to 500 mm.

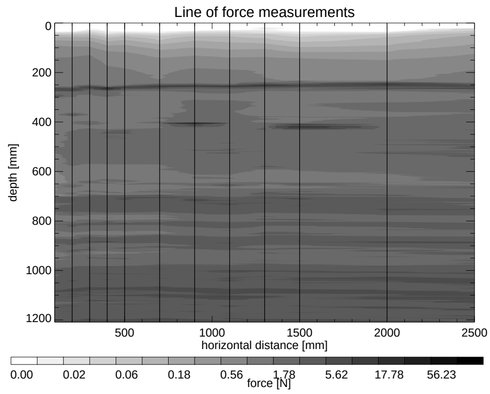

Fig. 13: Visualisation of two-dimensional spatial variability of a line of penetrometer snow profiles.

In Fig. 14, a three-dimensional view into a small, 25 x 15 cm snow block is illustrated. Measurements were taken in a regular 5 x 5 cm grid. Analysis of two- and three-dimensional spatial variability of the micro mechanical and micro textural properties of snowpacks can now be undertaken.

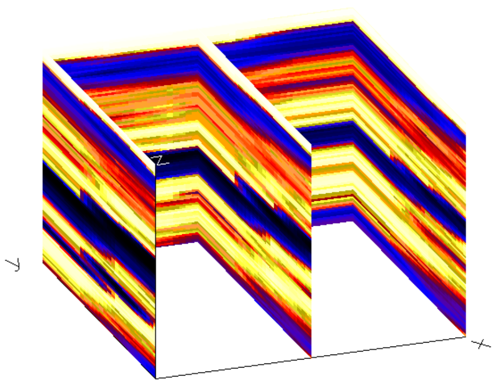

Fig. 14: Fence diagram of a small snow block and its stratigraphy.

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Chapter 4

Snow texture: a comparison of empirical versus simulated Texture Index for alpine snow

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Annals of Glaciology 32: 7-13, 2001

Abstract

The texture of snow has great impact on the mechanical and thermal properties of a snowpack. The Texture Index is a new concept to quantify the texture of snow at the micro and meso scale, a scale at which the properties of snow layers are responsible for the stability of a snowpack. It is calculated by taking the ratio of mean grain size (mm) of snow to density (kg m^{-3}) of snow in the respective layer. It combines a micro structural parameter with a meso structural parameter. An empirically derived Texture Index is compared with a simulated Texture Index for laboratory experimental snow and natural snow profiles. The Texture Index was calculated throughout the development of three temperature gradient snow experiments from laboratory measurements of grain size and density. In a one-dimensional finite element snowpack model (SNOWPACK), the development of the mean grain size and density of these snow samples were simulated. It can be shown that the measured Texture Index can be predicted by the modelled Texture Index. Furthermore, three natural snow profiles from the Weissfluhjoch test field are compared with their simulations in terms of Texture Index. The latter comparison reveals shortcomings in the measurement of hardness and grain morphological parameters. It shows also the dependency of a good physical model on quantitative process descriptions at the micro scale. The Texture Index can be estimated using the force signal of the SnowMicroPen, a new high-resolution snow micro penetrometer. It is suggested that more accurate hardness and grain morphological measurements combined with penetrometer measurements in the immediate vicinity of the snow profile will allow for a better verification of SNOWPACK simulations.

4.1 Introduction

Snow texture and its associated mechanical and thermal properties play a key role in avalanche research. Arons and Colbeck (1995) define the texture of snow in terms of both, micro structure and meso structure. Besides grain size and grain shape, the arrangement of the grains in a snow sample is included in this definition. This general definition takes the bonds between the grains into consideration. Micro scale features are grain shape and grain size. The meso structural features are more complex, they include the spatial arrangement of the grains and the intergranular connections. The idea behind the Texture Index (TI) is to combine the micro and meso scale features of a snow layer in order to express its textural properties (Schneebeli et al., 1999). The ratio of mean grain size and volumetric density decreases for smaller spherical structural elements. Furthermore, the TI is a direct index of the volume density of grain contacts, it encompasses the number and size of grain contacts. A low value of TI is interpreted as high volume density of grain contacts, which indicates high textural stability. A high value of Texture Index is interpreted as low volume density of grain contacts, which indicates low textural stability. Two natural snow profiles containing weak layers have been successfully classified by the TI (Schneebeli et al., 1999). In this approach, the TI is further calculated for laboratory snow samples taken by Baunach et al. (2000). Grain size and density are simulated for these samples by a one-dimensional snowpack model. A comparison of measured and simulated parameters shows good agreement. Hence, we can establish a link between an objective mechanical measurement of the snow cover and a numerical model. Since the empirically derived parameters also agree with the simulated ones, we also have a link to the classical snow cover investigation methods. Three natural classical snow profiles taken during the winter 1998/99 at the Weissfluhjoch test field are compared to their simulations. The snowpack was simulated using the data from the automatic snow station that is situated at the test field. The comparisons show that we cannot readily compare the TI calculated from a classical profile with the TI from its simulation.

4.2 The Texture Index of snow

The TI was formulated from a model conception of the bonding of snow in order to classify snow types from the micro penetrometer's force signal. The Texture Index ($m^4\ kg^{-1}$) is the fraction of mean grain size (mm) and density ($kg\ m^{-3}$):

$$TI = \frac{mean\ grain\ size}{density}$$

The idea behind the TI is that smaller spherical structural elements have a decreasing TI, i.e. a decreasing ratio between mean grain size and density. This ratio accounts for the connectedness or textural stability of the grains in that it is a direct index of the volume density of grain contacts. We use a dimensioned index because it describes a concrete geometric model. Moreover, the strength of the bonds themselves is responsible for the mechanical strength of the snow, however, the TI does not account for this. Basis for the formulation of the TI were 23 natural and artificial metamorphosed snow samples that were analysed in terms of mean grain size and density. The resulting Texture Indices could be correlated to the coefficient of variation (CV) of the micro penetrometer's force signal derived from these samples. The calculation of the equation for the least-squares line, the so-called linear regression, resulted in the following coefficients:

TI = 0.00145 + 0.00572CV

The coefficients are highly significant ($p<0.001$) and the coefficient of correlation r is 0.89 (Schneebeli et al., 1999). Based on this model the TI can be estimated with values of the coefficient of variation from the penetrometer's force signal. The coefficient of variation is calculated for intervals of 1000 force data values. Modelled values of TI below 0.003 indicate a high textural stability, values between 0.003 and 0.006 indicate a transition between textural stability and instability, and values greater than 0.006 indicate textural instability. With a combination of the TI and the median penetration force, snow profiles can be interpreted. The variation of the signal cannot be interpreted if the median is less than 0.03 N. These low values for the median correspond to the device travelling through air and it indicates a very soft layer or in fact air. Resonant frequencies start to superpose the variation of the force signal if the median is greater than 2.3 N. In this case, the variation of the force signal cannot be interpreted. These high values for the median correspond with very hard or ice layers in the snowpack, which are stable by nature.

4.3 Data

4.3.1 Experiments

During the winter 1998/99 experiments with artificially metamorphosed snow were conducted in the cold laboratory by Baunach et al. (2000) to study grain growth under different temperature gradients. Table 1 gives an overview of the initial state of the snow samples. Grain size and density measurements have been taken throughout the run of the experiments. We used the data from the experiments to compare the empirical TI from laboratory measurements with the TI derived from a SNOWPACK simulation of grain size and density.

Table 1: Summary of the initial state of the snow samples and boundary conditions of the snow experiments.

No.	Temp. gradient	Mean grain size	Density	Sample height	Number of days
	$K\ m^{-1}$	mm	$Kg\ m^{-3}$	m	
1	150	0.6	290	0.15	4
2	170	1.1	120	0.06	14
3	120	1.1	130	0.12	14

4.3.2 Experimental laboratory methods

To measure the volumetric density the content of a density shovel (6 cm x 5.5 cm x 3 cm) with 100 cm^{-3} volume is weighed. The density sample is carefully cut from the centre height of the snow block with the shovel, the net weight is determined and density in (kg m^{-3}) is calculated with an accuracy of ±5 kg m^{-3}. Afterwards, the density sample plus two horizontally adjacent samples are sieved to determine the grain size distribution (Sturm and Benson, 1997). The content of the three 100 cm^{-3} density shovels is sieved through a stack of five aluminium sieves (mesh openings from bottom up are 0.25 mm, 0.5 mm, 1 mm, 2 mm and 4 mm). For each grain size fraction the weight is determined with an accuracy of ±0.01 g and the mean grain size is calculated using the equation (Friedman and Sanders, 1983):

$$d_m = \frac{\sum_{j=1}^{L} M_j D_{j,j+1}}{100}$$

where d_m is the mean grain diameter for the sample, M_j is the weight fraction in the jth sieve, and $D_{j,j+1}$ is the average of the sieve mesh opening of the jth and j+1th sieves, with a total of L sieves. The error of the grain size d in one particular sieve is (Friedman and Sanders, 1983):

$$d = \pm \frac{D_{j+1} - D_j}{2n}$$

where n is the number of grains in the jth sieve. Very few crystals remained in the largest sieve and in the pan on the bottom of the stack. The respective grain diameters were set to 6 mm and to 0.15 mm. Assuming large *n* (3000 spherical ice grains of 1 mm^3 volume weigh about 2.75 g) and a normal distribution of the grain size in one particular sieve, the error around the mean approaches zero. The particular sieve receiving the largest weight dominates the accuracy of the mean grain size calculated from all sieves. Hence, by using large sample sizes we achieve a very small statistical error on the mean grain size determined by sieving. In the following comparisons of the laboratory experiments the mean grain size from sieve analysis is used. Baunach et al. (2000) found that the mean grain size from sieve analysis is ±10 % to the average diameter determined by image analysis using standard image processing software. Visual observation yields a very good approximation of the maximum diameter received from image analysis. However, the average grain size determined by the observers is closer to the minimum grain size determined by image analysis. To avoid a bias to extreme grain sizes in a sample analysed in the field, the observers of the Swiss Avalanche Warning Centre give a range of grain sizes instead of a single value when taking a classical snow profile.

4.3.3 Simulation

SNOWPACK is a one-dimensional, physical snowpack model based on the finite element method. It was developed for the Swiss National Avalanche Warning Service (Lehning and Bartelt, 1999; Lehning et al., 1999). The model is used to attend to the special requirements of avalanche warning in alpine Switzerland. One important task is the estimation of macro- and microscopic properties of individual layers and their development through time. Bulk density and grain size are included in these properties. The SNOWPACK model is used to calculate a simulation of the development of the laboratory snow experiments. In SNOWPACK the initial conditions are new snow height, grain size and new snow density (the case of dry snow is assumed). In the case of the laboratory experiments, the new snow height is equivalent to the height of the sample. Density of the new snow is equivalent to the measured initial density and in every experiment the same initial grain size of 0.6 mm is used. The grain size is adapted to this fixed value, so that the model can start calculating grain growth upon the total decomposition of the new snow dendrites (Baunach et al., 2000). Further initial conditions are given by the snow grain characteristics dendricity and sphericity (Lesaffre et al., 1998; Fierz and Baunach, 2000). The sphericity and dendricity for new snow are set to 0.5. The model adapts dendricity so that it is zero when grain growth commences. The boundary conditions in SNOWPACK are usually determined by the meteorological situation obtained from the automatic snow stations of the Swiss IMIS-network. Boundary conditions in the field for determining the conditions at the top of the snowpack are either heat-flux parameters (Neumann condition), i.e. short and long wave radiation as well as sensible and latent heat transfers or surface and bottom temperature of the snowpack (Dirichlet condition) (Lehning et al., 1999). In the simulation of the laboratory experiments the boundary conditions are set to the surface and bottom temperatures of the snow samples. With temperatures always below 0° C (case of dry snow) the Dirichlet condition is fulfilled and SNOWPACK uses the appropriate formulations. Temperatures are controlled during the experiments by thermocouples situated at the bottom, middle and on top of the snow sample in the metamorphosis box. After all initial and boundary conditions are defined we use SNOWPACK to calculate the development of grain size and density of our laboratory snow samples. These simulated values are then compared to the measurements taken throughout the experiment.

4.4 Comparison of laboratory experiments

To compare the simulated and measured TI from the laboratory snow, we take the ratio of mean grain size to density, both from the measurements and simulations described above. The simulation delivers half hour values for the two parameters. The measurements are taken in larger time steps. Fig. 1 shows the simulated versus measured TI for the three laboratory experiments. The symbols on the measured curves indicate the discrete point in time where measurements were taken.

4.4.1 Results

The comparison of simulated versus measured TI shows good agreement for all three experiments. Generally, the trend to an increasing TI, i.e. an increasing textural instability, is captured very well in all simulations. This trend is due to the fast formation of faceted snow under high temperature gradients. The developing depth hoar crystals grow in size while the number of bonds declines. Both experiments, number 2 and 3, show a significant underestimation of the measured values at the start of the experiment between day 0 and 1. This underestimated TI is due to an overestimation of the density at the start of the experiment. SNOWPACK delivers a settlement rate for new snow layers that is too small in the first two days after deposition. In the model calculations, the higher density values at the start were deliberately set as initial density to avoid underestimating the settlement in the beginning. With the TI we can easily detect this problem. It reoccurs in the simulation of the snow profiles and will be discussed in the next section.

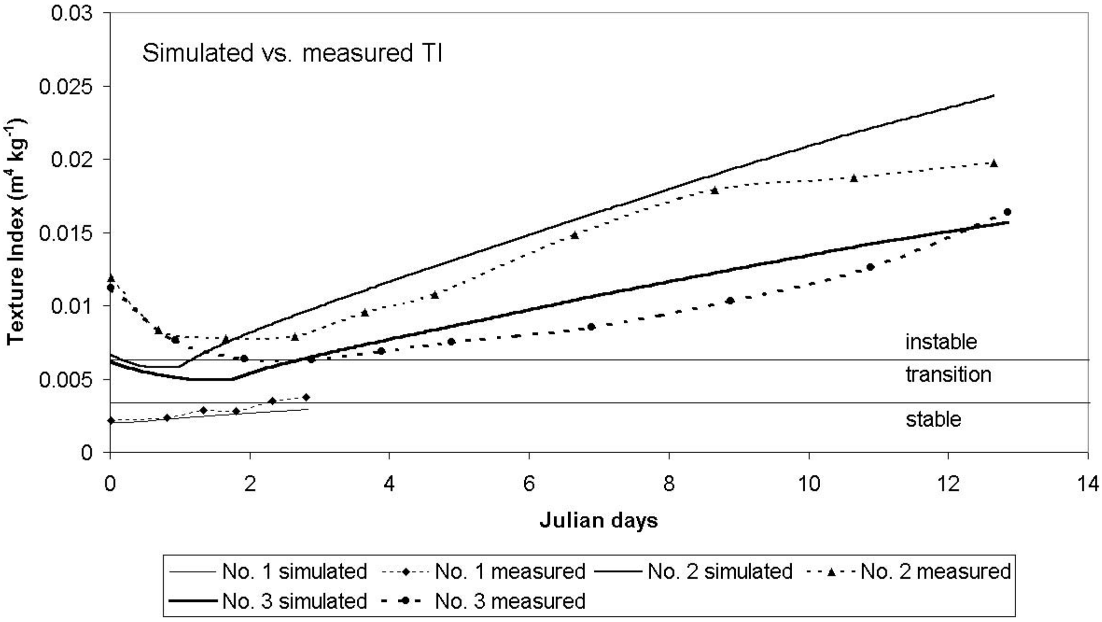

Fig. 1: TI for laboratory experiments 1-3.

4.5 Comparison of snow profiles

Snow profiles at the Weissfluhjoch test field are taken biweekly by the observers of the Swiss Avalanche Warning Centre. Grain size and density information are recorded and input on a profile sheet. An example profile sheet is shown in Fig. 2. The observer records ram hardness, temperature, stratigraphy, humidity, grain size and shape, hand hardness and density throughout the profile. The classification of grain shape and hand hardness is based on the Handbook for Observers (SLF, 1989). The observer takes density measurements between marker threads that are placed on the surface of the snowpack after each biweekly profile. Therefore the snow deposited over the previous fortnight becomes integrated in one density measurement. For grain size and grain shape the observer determines the individual layers for which these two parameters are recorded. Hence, measured density layers and grain size layers do not correspond. Grain size is measured by visual inspection (using a magnifying lens) of a large number of grains on a measuring plate. The observers do not record one single grain size for a layer but a range of grain sizes. The measured grain size value used in each of our comparisons is the mean of that given range. This practice is used to avoid a bias towards the extreme grain sizes in a sample and hence to reduce the error of the grain size measurements in the field. Three classical snow profiles taken by observers of the Swiss Avalanche Warning Service during the winter 1998/99 at the Weissfluhjoch test field are compared to their SNOWPACK simulations. The three profiles were taken on the following dates: 1 December 1998, 15 January 1999 and 16 February 1999. For all profiles the simulated and measured Texture Indices are compared and systematic deviations are depicted.

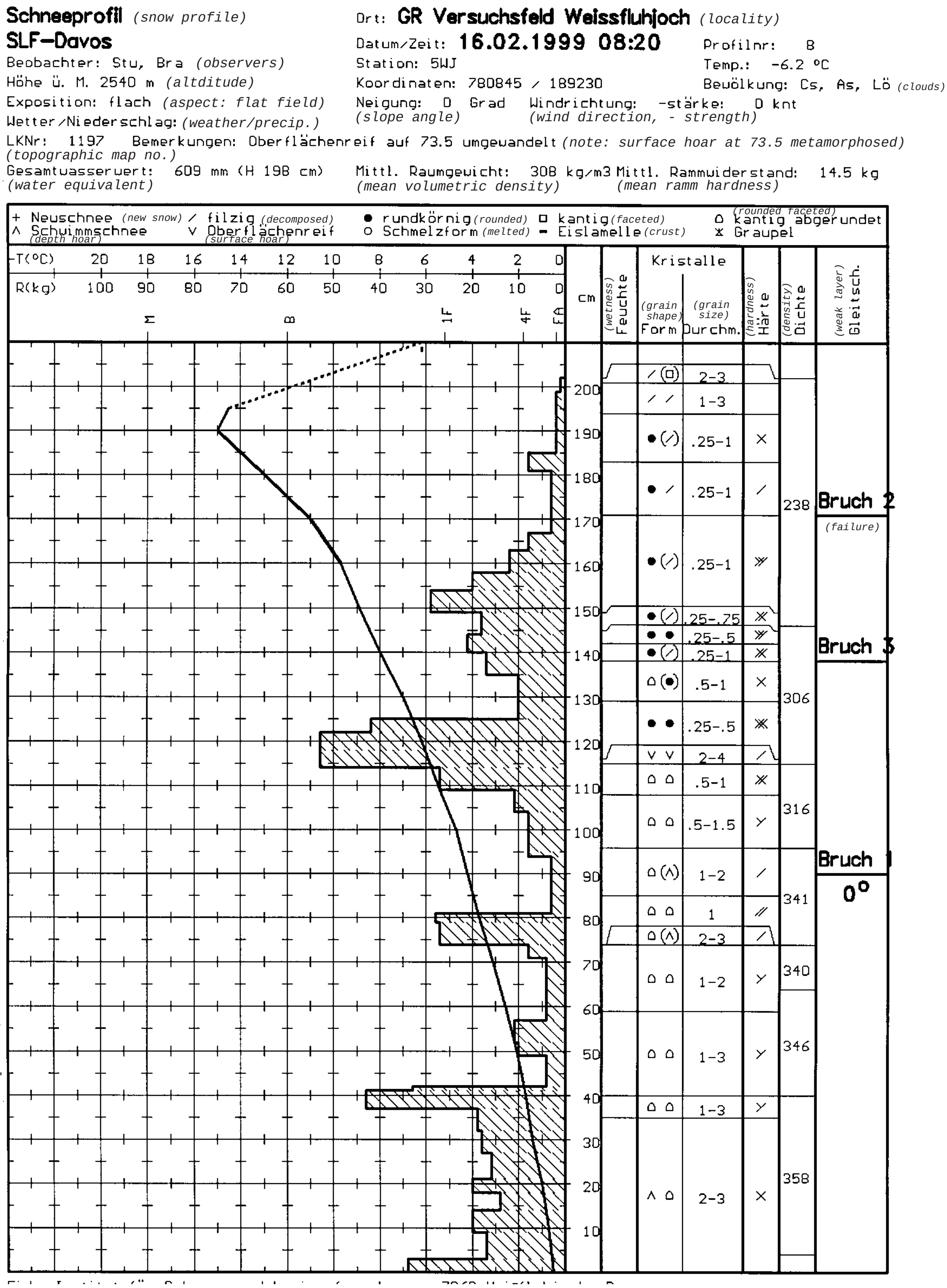

Fig 2: Classical snow profile.

4.5.1 Profile 1

The comparison for profile 1 is shown in Fig. 3. In this profile the simulated TI is generally too low compared to the measured TI. The high simulated TI (greater than 0.006) at the top of the profile represents a simulated surface hoar layer that was observed in the field but no density was measured. The high measured TI at the base of the profile is due to observed large grains that are not captured in the simulated profile. Fig. 4 shows the comparison of grain size and density for profile 1. In the following the deviations for grain size and density are described separately from bottom to top of the profile.

Grain size:

The simulated grain size is generally too small and in particular it is 1 mm too small at the bottom of the profile. In the lower part of the profile melt-freeze crusts and one ice lens were observed. Grain size was measured for the melt-freeze crusts but not for the ice lens. Ice lenses and crusts do not appear in the simulation. From the middle to top layers the simulated grain sizes match the measured well. The high value for grain size at the very top represents a simulated surface hoar layer that was also observed.

Density:

At the bottom of the profile the simulated density is significantly too high (about 100 kg m^{-3}). For melt-freeze crusts and ice lenses there is no measured density. Simulated and measured match well up to the middle of the profile where the simulated density is overestimated by 150 kg m^{-3}. Again there is good agreement up to the surface. The simulated surface hoar layer's density is taken to be 15 kg m^{-3} in SNOWPACK. The surface hoar layer was observed in the field but no density was measured.

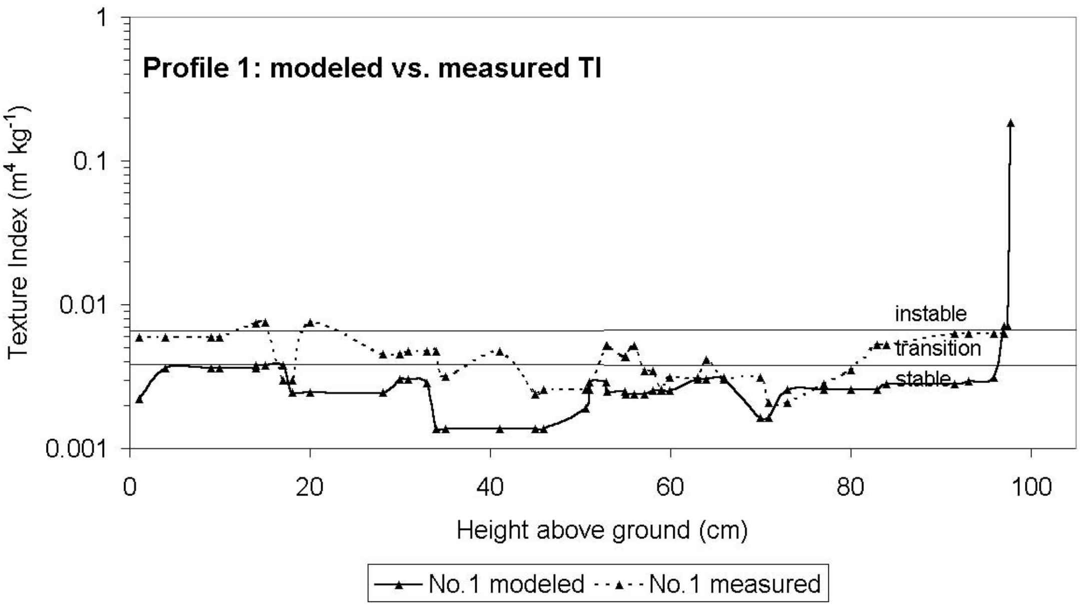

Fig. 3: TI comparison for profile 1.

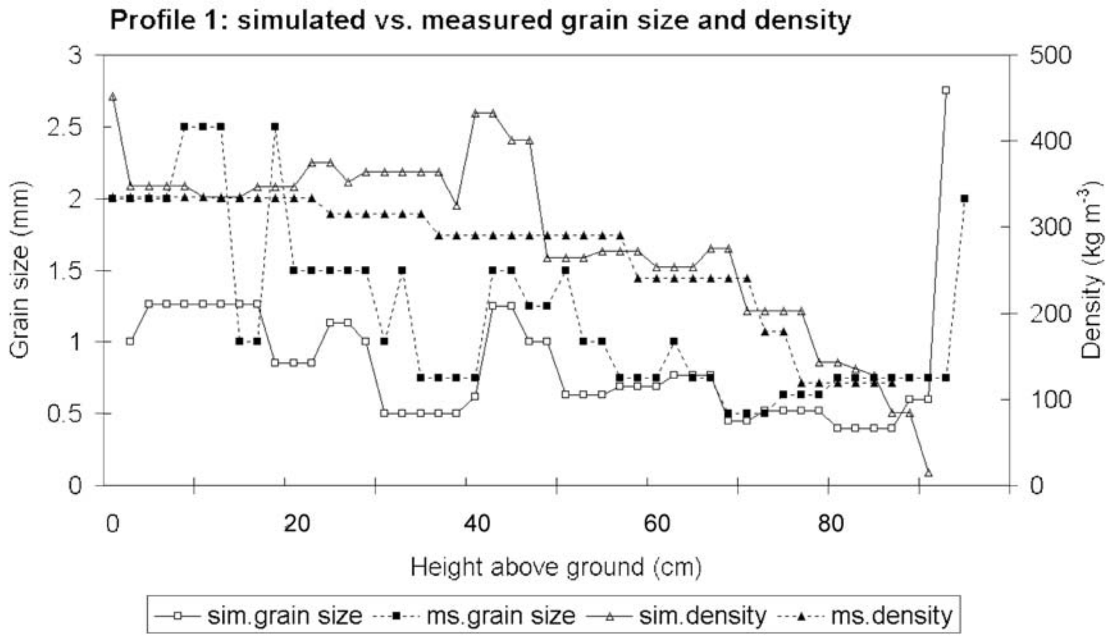

Fig. 4: Grain size and density for profile 1.

4.5.2 Profile 2

Fig. 5 is the TI comparison for profile 2. Since grain size for ice lenses was not measured in this profile, an empirical approximation of 5 mm is used. Therefore the measured TI is higher for the observed ice lens. Ice lenses do not appear at all in the simulation. The simulated TI is generally too low compared to the measured TI. The high measured TI at the base of the profile is due to observed large grains that are not captured by the simulated TI. In the upper half of the profile there are five TI-peaks that do not agree with the measured TI. These simulated peaks are well above a TI value of 0.006 and represent predicted buried surface hoar layers. These simulated buried surface hoar layers were not observed in the field. At the top of the profile a graupel layer was observed that does not appear in the simulation. Because of the relatively large grain size of the graupel, the measured TI is much higher than the simulated TI at the surface. Fig. 6 shows the comparison of grain size and density for profile 2.

Grain size:

At the bottom of the profile the simulated grain size is too small. Ice lenses and melt-freeze layers recorded by the observer are not simulated at all. The measured grain size of ice lenses is taken to be 5 mm. Two observed ice lenses occur in the lower half of the profile. In the upper half of the profile the simulated grain size shows 5 peaks, which represent buried surface hoar layers. The simulated grain size reaches 36 mm. Only in one instance a layer of buried surface hoar was measured in this profile. However, it differs in profile height (measured layer at 90 cm) and grain size (measured grain size 1-2 mm). Towards the surface of the profile the measured grain size increases steadily from 1 to 4 mm. The model does not capture this increase.

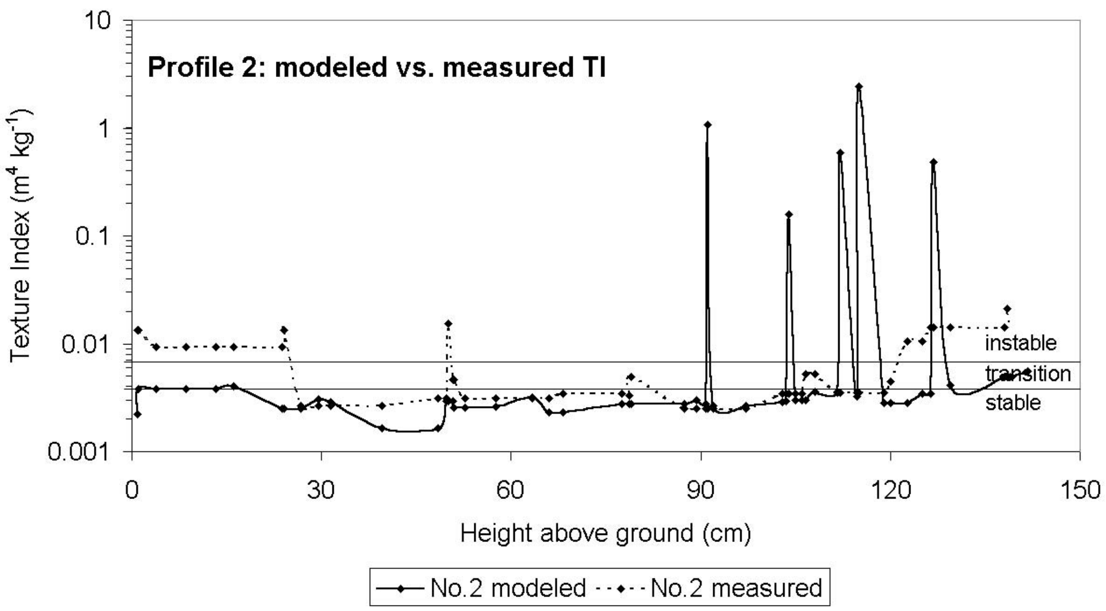

Fig. 5: TI comparison for profile 2.

Density:
At the very bottom of the profile the simulated density is about100 kg m^{-3} too high, as it was the case in profile 1. Throughout the profile the simulated density over- and underestimates the measured density by 50 and 100 kg m^{-3}. The buried surface hoar layers, with the very low modelled density of 15 kg m^{-3} are not measured in the snow profile. For the observed surface hoar layer density could not be measured. This layer does also not appear in the simulated density at this position. At the surface the simulated density agrees well with the measured density.

4.5.3 Profile 3
Fig. 7 is the TI comparison for profile 3. In the lower half of the profile the simulated TI is too low. The four simulated buried surface hoar layers were not observed in the field except for one layer at profile height 120 cm. This buried surface hoar layer was predicted very closely to the measured layer but it was simulated with a much higher TI. The predicted surface hoar layer just below the surface of the profile was not measured in the field. It appears as a simulated TI-peak. Fig. 8 shows the comparison of grain size and density for profile 3.

Grain size:
The higher measured TI at the bottom of the profile results from an observed depth hoar layer that does not appear in the simulated TI. The measured grain size in this layer exceeds the simulated by 1 mm. Three of the four simulated buried surface hoar layers in this profile were not observed. At a profile height of 120 cm there is one observed buried surface hoar layer in close proximity to the simulated one (1 cm). However, the simulated grain size is more than

double the measured (7 mm versus 3 mm). The simulated surface hoar layer just below the surface was not observed.

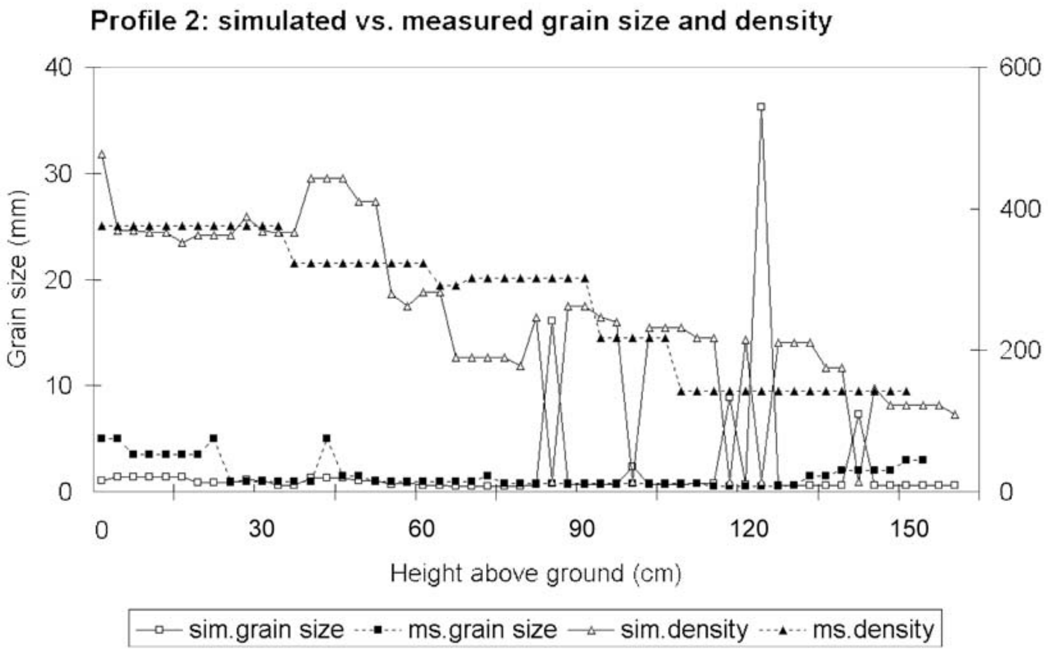

Fig. 6: Grain size and density for profile 2.

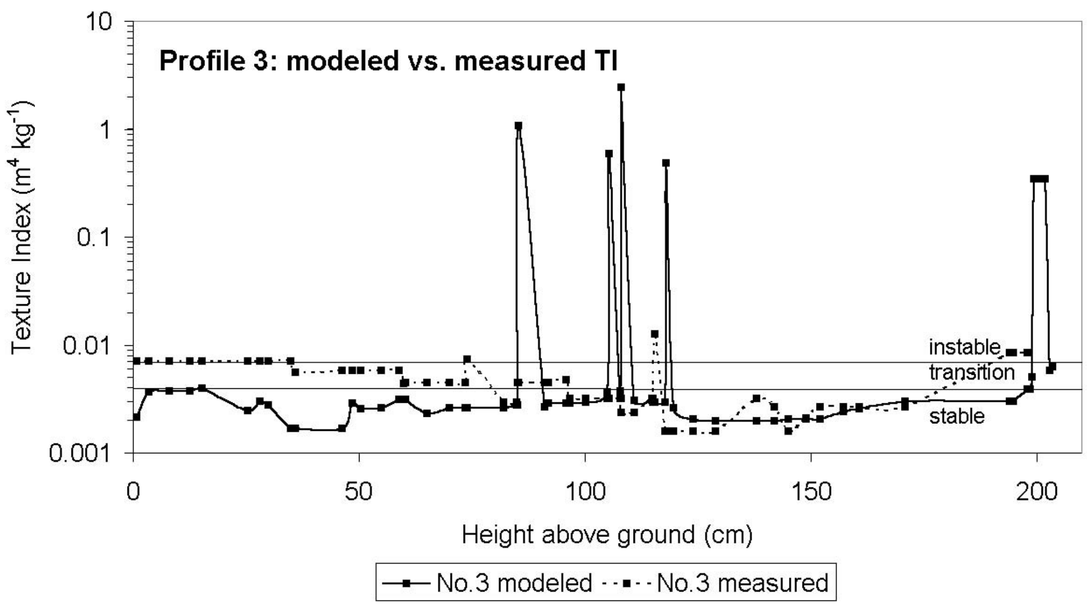

Fig. 7: TI comparison for profile 3.

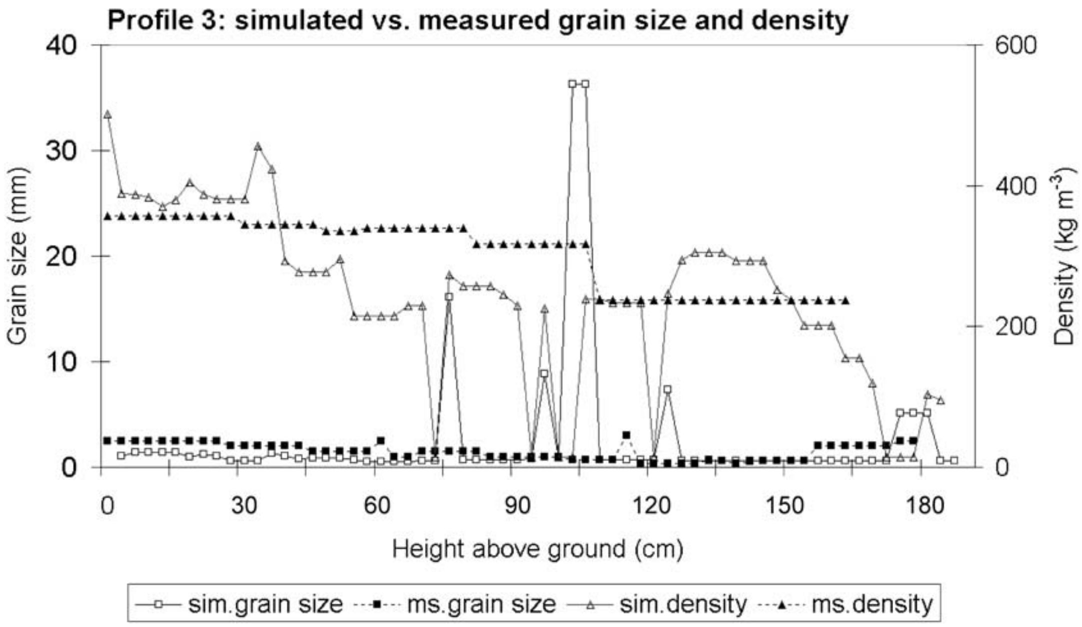

Fig. 8: Grain size and density for profile 3.

Density:
The simulated density at the bottom of the profile is 150 kg m^{-3} higher than the measured density. At 60 cm profile height, the simulated density drops 100 kg m^{-3} below the measured density. By definition, the simulated density of the buried surface hoar layers is taken to be 15 kg m^{-3} in SNOWPACK. In case of the observed buried surface hoar layer no density was measured because the layer was too fine for a measurement. The measured density here is the integrated density for the larger layer surrounding the buried surface hoar layer. Increased grain size was measured in the surface layer and the layer just below the surface. However, density measurements are not available for these layers.

4.5.4 Results

- The simulated grain size at the bottom of the profiles is too small.

- Ice lenses and crusts or melt-freeze layers do not appear in the SNOWPACK simulation. They are qualitatively described in the measured profile, and only in a few cases the grain size could be measured. Density of ice lenses and crusts cannot be measured in the field.

- SNOWPACK calculates surface hoar as soon as the conditions fulfil the minimum requirements for its development. This guarantees that surface hoar is simulated in any case when it actually occurs, but often the simulated surface hoar is not observed in the field. This was the case at the surface of two profiles. Instead of surface hoar, the observer recorded a graupel layer, and in the other case decomposing fragmented particles at the surface were recorded.

- Consequently, the buried surface hoar layers are too many in the simulated profiles. Out of all simulated buried surface hoar layers only two were actually observed. For these measured buried surface hoar layers the observed grain size was much smaller than the simulated grain size.

- The low density of the simulated buried surface hoar layers has not been measured.

- The simulated density at the bottom of the profiles is too high.

4.6 Discussion

The comparison of the TI for the three laboratory experiments shows good agreement. This was in part expected because these experiments were conducted to improve the model specifications, particularly the laws of grain growth under strong temperature gradients.
The comparison of the snow profiles shows a number of systematic deviations between simulated and measured TI. The deviations of the absolute profile height are due to spatial variability. In case of the missing simulated crusts and ice lenses, SNOWPACK must be adapted in the future to display the melt-freeze features. The density of crusts and ice lenses cannot be easily measured so far in the field. However, an empirical value (about 500 kg m^{-3}) obtained from image analysis is better suited to account for the actual density discontinuity. Often these abrupt discontinuities are weak interfaces within a snowpack.
Another major deviation appears with the buried surface hoar layers. Buried surface hoar layers are displayed in the simulation up to 5 layers within one profile. For visualisation purposes, if the number of layers exceeds 5, the least pronounced surface hoar layer drops to the background of the calculations and is no longer displayed in the simulation. Also by definition, the simulated grain size of buried surface hoar layers is taken to be equal to the height of this layer and the simulated density is taken to be 15 kg m^{-3} in SNOWPACK. Upon completion, these defined values will be adapted to the results of an extended surface hoar study at the SLF. According to Föhn (personal communication) the following values are most likely: surface hoar at the surface has a mean density of 100 kg m^{-3}; buried surface hoar has a mean density between 100 and 250 kg m^{-3}; grain size is highly variable, ranging from mm to cm. SNOWPACK calculates more surface hoar layers than are observed. Once simulated at the surface, these layers remain constant as buried surface hoar layers in the simulated profile. Field observations, however, show that buried surface hoar layers can be destructively metamorphosed. Because the density measurements in the field are to crude to pick up these fine layers and the grain size observations are highly subjective it is impossible at the moment to verify the existence of all the buried surface hoar layers that appear in the model.
The settlement at the near surface is simulated well, however in the middle of the profile settlement is generally to low and at the bottom to fast. This explains that the simulated density is significantly too high in the upper part of the profile and significantly too low in the lower part of the profile. Also, new snow settles too slowly in the model.
The model simulates grain growth under high temperature gradients only after complete decomposition of the dendrites. However, such metamorphism has also been observed in the field with partially decomposed dendrites.

It is likely that the classical grain size and density measurements are too crude to resolve fine layers within the snowpack. Grain size and density measurements need the same resolution to make them readily comparable to the simulation. With the methods applied so far, we cannot verify fine layers that are simulated. Highly resolved measurements are necessary to capture the heterogeneity of the layers and within the layers of a natural snowpack. The TI can be derived from the force signal of a snow micro penetrometer. Therefore higher resolved field measurements plus penetrometer measurements in the immediate vicinity of the snow profile will allow for a better verification of SNOWPACK simulations.
The TI can be used to partially interpret the snow stability of layers in a snowpack. The number of contacts per unit volume is an indication of snow stability. Moreover, the strength of the bonds themselves is responsible for the mechanical strength of the snow, which is not accounted for by the TI. However, a combined analysis of TI and median penetration force allows interpreting the micro texture and bond strength. With stereological methods we cannot properly measure bond size and bond strength. Methods for the three-dimensional reconstruction of snow are now being further developed to approach this problem. The TI was chosen for the comparison for it is thought to be the lowest common denominator in the approach to micro textural characterisation of snow in the field.

4.7 Conclusion

The TI is a useful measure to partially interpret the mechanical stability of snow because it can be determined easily in the field by classical snow cover investigation methods. For laboratory snow samples, a numerical snowpack model can predict the TI. However, systematic deviations between field measurements and simulations have appeared in the comparison of three natural snow profiles. The discussion of the deviations reveals that improvements in the model formulations as well as improvements in field methods are necessary to make them readily comparable. Highly accurate measurements are necessary to capture the heterogeneity of the snow properties of a layered natural snowpack. Especially thin layers cannot be sufficiently quantified by the applied field methods. Yet they have a major influence on mechanical and thermal processes and on the stability of a snowpack. The TI derived from the force signal of a snow micro penetrometer can easily detect these thin layers. More accurate classical measurements combined with penetrometer measurements in the immediate vicinity of the snow profile will allow for a better verification of SNOWPACK simulations. The TI and the median penetration force obtained from the penetrometer give a good and fast indication of weak layers in a snowpack. The information obtained from the penetrometer profile could be further used to apply additional mechanical tests (shear frame, rutschblock) on an identified weak layer. Highly resolved field measurements improve the profile interpretation, the avalanche warning and avalanche prediction. They also allow for a better model verification, which leads to improved simulations, which can be used in avalanche warning and prediction.

4.8 Acknowledgement

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Chapter 5

Soil thermal conditions on a ski-slope and below a natural snow cover in a sub-alpine ski resort

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Abstract

Earlier studies have indicated that the soil on an intensively prepared ski-slope may be subjected to more pronounced cooling than the soil below a natural snowpack. We analysed the thermal impacts of ski-slope preparation in a sub-alpine ski resort in central Switzerland at an altitude of 1100-1200 m a.s.l. where artificial snow is produced. Soil temperature measurements were carried out on ski-slopes and off-piste during one winter season (1999/2000). In addition, a numerical soil-snow-atmosphere transfer model with a new-implemented option to simulate the snowpack development on a groomed ski-slope was run for one of the ski-slope and off-piste sites. The model proved to be an appropriate tool to simulate changing physical snow properties on and beside ski-slopes. For the soil temperatures, the considerable differences in the snow thermal properties between ski-slope and off-piste showed to be less important than the fact that air temperature was close to the freezing point for most of the winter season. Hence, the differences in soil temperatures between the ski-slope and the natural snowpack outside the ski-slope were not significant according to the simulations, which was supported by the soil temperature measurements.

5.1 Background

In the Alps, many ski resorts are situated at relatively low altitudes of 1000 to 1500 m a.s.l. Here, a changing climate and, as a consequence of that, an expected decrease of the snow cover will have a growing impact on the future of these ski resorts, since snow conditions sufficient for skiing will be more and more uncertain at this altitude (Bürki and Elsasser, 2000). On the other hand, the continuously increasing demand by ski tourists has prolonged the skiing seasons from the end of November until mid of April. As a consequence, the production of artificial snow and ski-slope preparation has gradually been intensified (Mosimann, 1991; Mosimann, 1998). At the same time the question has been raised what effect such intense artificial snow generation and ski-slope grooming would have on the soil ecology. Since the end of the 1980s, several aspects of the interrelation between ski-slopes and the soil ecosystem have been studied, such as the fertilising effect of artificial snow on the ground vegetation (Cernusca, 1984; Kammer and Hegg, 1990) and the damage to grown vegetation and soil structure from machine-grading (Titus and Tsuyuzaki, 1999).

With regard to the soil physical impact of ski-slopes, there are two major issues that have been the subject of research studies: (a) the availability or lack of oxygen below the tightly pressed snowpack (Newesely and others, 1994), and (b) the change in soil thermal conditions due to a marked alteration of the thermal properties of the snowpack. Both impacts are likely to influence the growth and composition of plant species for ski-slopes over the long term (Cernusca and others, 1989). Although artificial snow has been produced for more than 20 years, and intensive snow grooming has been carried out for a considerably longer time, the issue of thermal conditions below a ski-slope has only been addressed in a small number of research studies. Mellini (1996) investigated the impact of ski-slope preparation on the distribution of permafrost, and Newesely (1997) measured soil temperatures on ski-slopes and below natural snow at four ski resorts in the Austrian Alps. Both studies found a more pronounced cooling on the ski-slope and a delayed thawing in spring. On the other hand, to our knowledge no study has been published where the thermal conditions on artificial ski-slopes have been studied with a numerical simulation model. Many such models have been developed for the simulation of natural snowpack development with respect to hydrological or micro-meteorological purposes, as well as for avalanche risk assessment. Their complexity ranges from simple one-layer snowpack models (e.g. Tarboton and Luce, 1996) to very detailed multiple-layer snow models (e.g. Brun and others, 1989; Jordan, 1991). Usually, such models are driven by standard meteorological measurements and describe the most important processes regulating accumulation, settling and melting, and in many cases the heat and water fluxes from/to the soil. However, when we attempt to simulate these processes for an intensively prepared ski-slope, it is obvious that we cannot base our simulations on meteorological records alone but we must also include information about the actual artificial snow production, and the ski-slope grooming. This is, however, a very difficult task because information about the rate of snow production is often lacking. And even if it were available, it would be very difficult to determine how this artificial snow was distributed over the ski slopes. Furthermore, the snow compaction due to the grooming by piste machines is very difficult to estimate. It depends on the state of the snow at the moment of preparation and the grooming intensity, but few or no measurements that would allow us to quantify such effects have been published.

Consequently, the most promising solution seems to be to use repeated observations of the snow depth and snow density to drive the simulation if the primary goal is to calculate the heat (and water) fluxes in and below the ski-slope snowpack. Such a strategy was also proposed by Lehning and others (1999) for natural snow covers where no precipitation measurements are available.

A further difficulty with regard to the simulation of a ski-slope concerns the radiation balance on the surface. It has to be assumed that frequent grooming leads to a considerably altered surface albedo compared with a natural snowpack. Unfortunately, only very few measurements have been reported concerning this issue (Nakamura and others, 1998). A suitable model for heat and water transfer of a ski-slope also needs to include a dynamic simulation of the processes in the soil based on soil physical properties - in particular soil freezing and thawing. One of the numerical SVAT2-models with a detailed description of soil freezing and thawing is the COUP-model3 (Jansson, 1998), originally developed with a major focus on heat and water fluxes in the frozen and unfrozen soil and later complemented with detailed descriptions of the processes at the soil surface, such as plant water uptake, surface heat exchange or snowpack development. The COUP-model has frequently been applied to winter locations (e.g. Johnsson and Lundin, 1991; Stadler and others, 1997; Jaesche, 1999; Stähli and others, 1999) and thus has become a suitable tool for studying practical issues related to snow and frost. Recent model development concerned the energy balance above a shallow snow cover and frozen soil, especially with respect to the turbulent heat fluxes (Gustafsson and others, 2001). In this study, we will for the first time apply the COUP-model to a ski-slope. For this purpose, an option has been added which allows a step-wise or continuous adjustment of the snow depth to observed snow depth measurements on the ski-slope.

The aim of this paper was

a) to test the new model approach for simulating heat and water dynamics in and below a ski-slope snowpack,
b) to determine differences in snow thermal properties between a ski-slope and a natural, off-piste snowpack, and
c) to evaluate how these different snowpack properties influence the soil temperatures.

5.2 Materials and method

5.2.1 Site description

The study was carried out on two slopes of the high-montane ski resort Brunni-Haggenegg (1100-1200 m a.s.l.) located in the Alptal-valley in Central Switzerland. The ski-slopes on the southeast exposure (Haggenegg) have artificial snow production and are intensively used by skiers, while the ski-slopes on the northwestern exposure (Erlenhöhe) contain natural snow only and are rather moderately skied and groomed. The soils on both slopes are characterised by a shallow groundwater level and, thus, very wet conditions (FAO classification: Dystric Gleysol). The bedrock is calcareous sediment (Flysch). Winter seasons in this high-montane region of Central Switzerland have typically rather mild air temperatures and high precipitation.

5.2.2 Weather conditions during the experimental winter

Compared with the 20 most recent years the experimental winter 1999/2000 was somewhat colder and higher in precipitation than average. The mean air temperature from November to March was only slightly below freezing at –0.7°C (long term mean November to March: 0.0°C). The lowest air temperatures at night were around –12°C. The accumulated precipitation in the same period was 1121 mm, compared to the long term mean of 821 mm. Grooming of the ski-slopes started in mid-November at Haggenegg and at the beginning of December at Erlenhöhe. The artificial snow production on the Haggenegg slope started mid-November as well. Snow was produced intensively in early winter, and the production was

[2] soil-vegetation-atmosphere transfer

[3] former name: the SOIL-model

stopped in January when enough snow had accumulated. No artificial snow was produced on the Erlenhöhe slope.

5.2.3 Measurements

During the winter 1999-2000 we continuously measured soil temperature profiles at two locations of the Haggenegg slope and the Erlenhöhe slope, one location on the groomed ski-slope and one location outside the groomed ski-slope where the snow cover was undisturbed (Table 1). For this purpose, thermistors (Yellow Springs Instrument Co.) were installed in the previous autumn at depths of 3, 20, 50 and 100 cm at each location. The accuracy of these sensors was ±0.5°C. In addition, a snow temperature profile was measured at Erlenhöhe outside the groomed ski-slope using thermocouples that were mounted vertically from the soil surface to heights of 3, 20, 50 and 100 cm above the soil surface. These probes were gradually covered by the snow. The uppermost sensor was snow covered from the beginning of January to mid March. Standard meteorological measurements were run every 10 minutes at Erlenhöhe, including air temperature, precipitation, global radiation, relative humidity and wind speed. Global radiation was also recorded at the Haggenegg slope where shading from the surrounding terrain was significantly different.

Table 1. Characteristics of the study sites.

Site	Artificial snow	Grooming intensity	Elevation (m a.s.l.)	Slope ( )	Aspect
Erlenhöhe, ski-slope	No	Low	1210	0.1-0.2	W
Erlenhöhe, off-piste	No	-	1210	0.25	W
Haggenegg, ski-slope	Yes	High	1120	0.15	E
Haggenegg, off-piste	No	-	1130	0.1-0.15	SE

Snow depth and density were manually recorded at all sites approximately once per week. The snow depth was measured with an avalanche probe (5 replications per site within 1 m^2), and the snow density with a steel cylinder having a diameter of 10 cm (3 replications). After mid December the snow on the Haggenegg ski-slope was too hard to allow manual snow density measurements (> 550 kg m^{-3}). At the end of the skiing season, we made an additional measurement of the snow density on the ski-slope by excavating a sample of known volume. Snow water equivalent was calculated from the snow depth and density.

In addition to the seasonal measurements, we made a detailed survey of snow properties on one selected day of the winter, on 12 January, a cold night with air temperatures below–8° C. The main purpose was to show how properties regulating the heat flux in the snowpack differed between the ski-slope and the off-piste sites. At all four locations we made the following measurements: A density profile was determined gravimetrically by sampling the snowpack from the top to the bottom with a steel cylinder having a volume of 0.001 m^3 and a height of 10 cm. The thermal conductivity was measured for the uppermost layers of the snowpack using the commercially available ISOMET 104 (ISOMET 104 User's Guide). This instrument is a microprocessor controlled portable instrument for the direct measurement of thermophysical properties of porous materials based on the hot wire method. A constant heat impulse is led into the snow via the needle of the probe and the temperature response of the snow is measured (Sturm and Johnson, 1992). The needle probe had a length of 10 cm, the actual temperature measurement occurred in the upper 7 cm of the needle. The thermal conductivity could not be determined in the lower part of the profile because the ISOMET-method gives reliable results only at snow temperatures below -2°C. In the lower part or the snow profile, however, the temperature was close to 0°C. Snow hardness was measured using the SnowMicroPen, a high-resolution penetrometer (Schneebeli and Johnson, 1998). Measurements were taken perpendicular to the snow surface down to a depth of 25 cm. The penetrometer was driven at a constant speed (20 mm s^{-1}) into the snow and recorded the penetration force exerted on the (5 mm diameter) measuring tip at a resolution of every 4 μm. This high resolution allows interpreting the snow hardness and texture in the mm-range. However, in very dense snow, such as on ski-slopes, for technical reasons only the hardness can be interpreted from the penetrometer signal (Schneebeli and others, 1999). For each location, we ran three replications and determined the average signal.

5.2.4 Model description

For the data analysis we used the one-dimensional numerical SVAT-model COUP (Jansson, 1998), extended with the snow model code of Jordan (1991). The COUP-model calculates combined heat and water fluxes in a layered soil profile and the overlying snowpack. The two basic assumptions for the soil are Richards' equation (Richards, 1931) for water flow, and the Fourier's law for heat flow. The boundary conditions are determined by formulations of interception, plant water uptake, drainage flow, percolation, surface heat balance and geothermal heat flux. A complete model description is given in Jansson and Karlberg (2001) and on the URL-site: http://www.lwr.kth.se/Coup.htm.

5.2.4.1 Input data

Three classes of input data are needed: (a) Driving variables, which are the climatic data governing the dynamics of the fluxes. Usually, they are provided as daily or hourly mean values. (b) Initial conditions defining the water and heat content of the soil layers at the start of the simulation. (c) Model parameters, which are the constants needed to express relevant properties for the different processes in the model.

The model requires the input of the following meteorological variables: precipitation, air temperature, relative air humidity, wind speed, net radiation or global short-wave radiation. All precipitation is assumed to fall as rain for air temperatures above T_{max} (usually set to +2° C) and snow for air temperatures below T_{min} (usually set to 0° C). Between these limits the transition from snow to rain varies exponentially.

5.2.4.2 Surface energy balance approach

The surface temperature, either for the snowpack or the bare soil, is calculated from the surface energy balance applying the law of conservation of energy,

$$R_{ns} = LE + H + q_{prec} + q_h \qquad (1)$$

where R_{ns} is the available net radiation at the surface, LE is the latent heat flow to the air, H is the sensible heat flow to the air, q_{prec} is the heat input from precipitation, and q_h is the heat

flow to the soil or the snowpack respectively. More details on the surface energy balance calculation in the COUP-model, especially with regard to the snow cover, are given in Gustafsson and others (2001).
For snow free situations, the albedo depends on the wetness of the topsoil as well as on the vegetation, whereas in the case of a snow cover the albedo, α_{snow}, is calculated following the approach of Plüss (1997):

$$\alpha_{snow} = \alpha_{min} + a_1 e^{a_2 n_d + a_3 T_{acc}} \qquad (2)$$

where a_1, a_2, a_3 are empirical parameters, α_{min} is the minimum albedo of the snow surface, n_d is the number of days since the last significant snowfall, and T_{acc} is the sum of positive daily mean air temperature since the last snowfall.

5.2.4.3 Multiple layer snowpack model

The entire snowpack is considered to consist of layers of varying density and thickness. A perfectly frozen fresh snow layer is assumed to have a minimum density of ρ_{smin}. Settling of the snowpack is a function of destructive metamorphism, overburden pressure, and snow melt:

$$CR = -\frac{1}{\Delta z}\frac{\partial \Delta z}{\partial t} = CR_{metam} + CR_{overb} + CR_{melt} \qquad (3)$$

where CR is the compaction rate (s^{-1}) and z (m) is the layer thickness. The thermal conductivity of the snowpack, $k_{h\ snow}$,, is related to its density according to

$$k_{h\ snow} = s_k \rho^2_{snow} \qquad (4)$$

where s_k is an empirical parameter.

5.2.4.4 New option for simulating a ski-slope snowpack

A new option was added to the model enabling the simulated snow depth to match observed snow depth measurements when these are available. This adjustment of the snow depth may either be discrete reflecting single events where snow was added or compacted (for example, from snow-guns or compacted with piste machines), or it can be continuous, representing diffuse or continuous snow redistribution processes such as snow drift by wind. In both cases, the natural snowfall is not excluded at any time, but still treated as explained above. If there is a deficit of simulated snow depth compared to the measurement, a snow water equivalent (SWE) corresponding to the snow depth deficit and the actual snow density will be added. The opposite applies when a snow depth surplus was simulated. The snow density after an adjustment remains thus unchanged.
As a consequence of periodical snow grooming, the surface properties of a ski-slope are expected to alter much more rapidly (i.e. after each action) than those of a natural snowpack which slowly change by the natural processes of snow melting and metamorphism. This applies above all to the albedo. However, there was no basis for a new formulation of the albedo since - to our knowledge - no studies have been reported where the change in albedo due to snow grooming has been quantified. Consequently, we use the same albedo function

(Eq. 2) for both natural and groomed snow covers, but change the parameter a_2 governing the time-dependent decline of the albedo so that the minimum value of α_{snow} is reached after some few hours.

5.2.4.5 Thermal dynamics in the soil

Below a snow cover, the soil surface temperature, T_{soils},, is given by assuming steady state heat flow between soil and snowpack:

$$T_{soils} = \frac{T_{layer1} + aT_{snows}}{1+a} \qquad (5)$$

where T_{layer1} is the temperature of the uppermost soil layer, T_{snows} the snow surface temperature, and a is a weighting factor given by:

$$a = \frac{k_{h\ snow}\frac{\Delta z_{layer1}}{2}}{k_{h\ layer1}\Delta z_{snow}} \qquad (6)$$

where Δz denotes thickness and k_h thermal conductivity. If the amount of liquid water in the snowpack exceeds a constant threshold, T_{soils} is set equal to 0° C.

The thermal conductivity for unfrozen mineral soil, $k_{h(min)}$,, is calculated with the empirical function suggested by Kersten (1949):

$$k_{h(min)} = 0.143(b_1 log\frac{\theta_w}{\rho_s} + b_2)10^{b_3\rho_s} \qquad (7)$$

where b_1, b_2, b_3 are constants, ρ_s is the dry bulk soil density, and θ_w is the water content.

When the mineral soil is frozen, the Kersten equation is modified with a higher order term to account for the influence of ice:

$$k_{h(frozen)} = c_1 10^{c_2\rho_s} + c_3\frac{\theta_w}{\rho_s}10^{c_4\rho_s} \qquad (8)$$

where c_1 to c_4 are empirical constants. The assumption is made that all water at the temperature T_f is frozen except for a residual unfrozen amount, $\theta_{uf,}$,

$$\theta_{uf} = d_1\theta_{wilt} \qquad (9)$$

where d_1 is a parameter and θ_{wilt} is the water content at the wilting point. For temperatures between 0 °C and T_f a weighted thermal conductivity is used according to the mass ratio of frozen water to total amount of water. The soil heat capacity equals the sum of heat capacities of the respective soil constituents multiplied by their volumetric fraction.

Treatment of frost in the soil is based on a function for freezing point depression and on an analogy between processes of freezing-thawing and drying-wetting, i. e. the liquid-ice interface is considered to be equal to the liquid-air interface. Thus, unfrozen water below zero is associated with a matric potential and an unsaturated conductivity. Freezing gives rise to a potential gradient, which in turn forces a water flow according to the prevailing conductivity. This causes a capillary rise of water towards the frost zone and it also allows drainage of snow melt through the frost zone when frozen soil temperatures are close to 0°C. A full description of this part is found in Stähli and Stadler (1997).

5.2.5 Model application and parameter setting

The model was applied to the Haggenegg slope for which both the off-piste site and the ski-slope were simulated. Hourly mean values of air temperature, relative humidity, wind speed and precipitation from Erlenhöhe, and global radiation from Haggenegg were inputs to the model. The parameterisation for the surface heat exchange was mainly based on the work of Gustafsson and others (2001).

The snow property data from the special survey on 12 January were used to estimate the coefficient s_k in the thermal conductivity function (Eq. 4).

The parameterisation of the soil physical properties (Table 2) was based on the measurements of Stadler and others (1998) and Stähli and Stadler (1997) who determined t4he pF-curve, k_{sat},, as well as the freezing characteristic curve of the Alptal-loam in the laboratory.

Initial soil temperatures were set according to the measurements, and a nearly saturated soil profile was assumed for the Alptal-loam, which corresponded to our observations.

Table 2. Parameterisation of the soil physical properties.

Horizon Depth interval	A_a-$G_{o,r}$ 0 - 10 cm	$G_{o,r}$ 10 - 45 cm	G_r 45 cm -
pore size distr. index [λ]	0.06	0.06	0.06
air entry pressure [ψ_a]	10 cm	10 cm	10 cm
porosity [θ_s]	0.73 m^3 m^{-3}	0.67 m^3 m^{-3}	0.61 m^3 m^{-3}
wilting point [θ_{wilt}]	0.28 m^3 m^{-3}	0.25 m^3 m^{-3}	0.21 m^3 m^{-3}
sat. hydr. cond. [k_{sat}]	6.22 103 mm d^{-1}	0.18 10^3 mm d^{-1}	0.49 10^3 mm d^{-1}
d_1 [Eq. 9]		0.5	
d_2 [Eq. 17 in Stähli and Stadler, 1998]		140	

5.3 Results

5.3.1 Seasonal development of the snowpack

The seasonal development of snow depth, mean snow density and snow water equivalent at the four locations are shown in Fig. 1. At the off-piste sites, we measured a peak snow accumulation of 40-70 cm already in the beginning of December, followed by an inter-mediate melting which led to an almost complete depletion on the south-east exposed slope Haggenegg. A second intense snowfall at the beginning of January produced a snowpack with a total depth of 60 to 120 cm outside the ski-slopes, corresponding to a snow water equivalent of up to 350 mm. On the ski-slope Haggenegg, the snow depth was

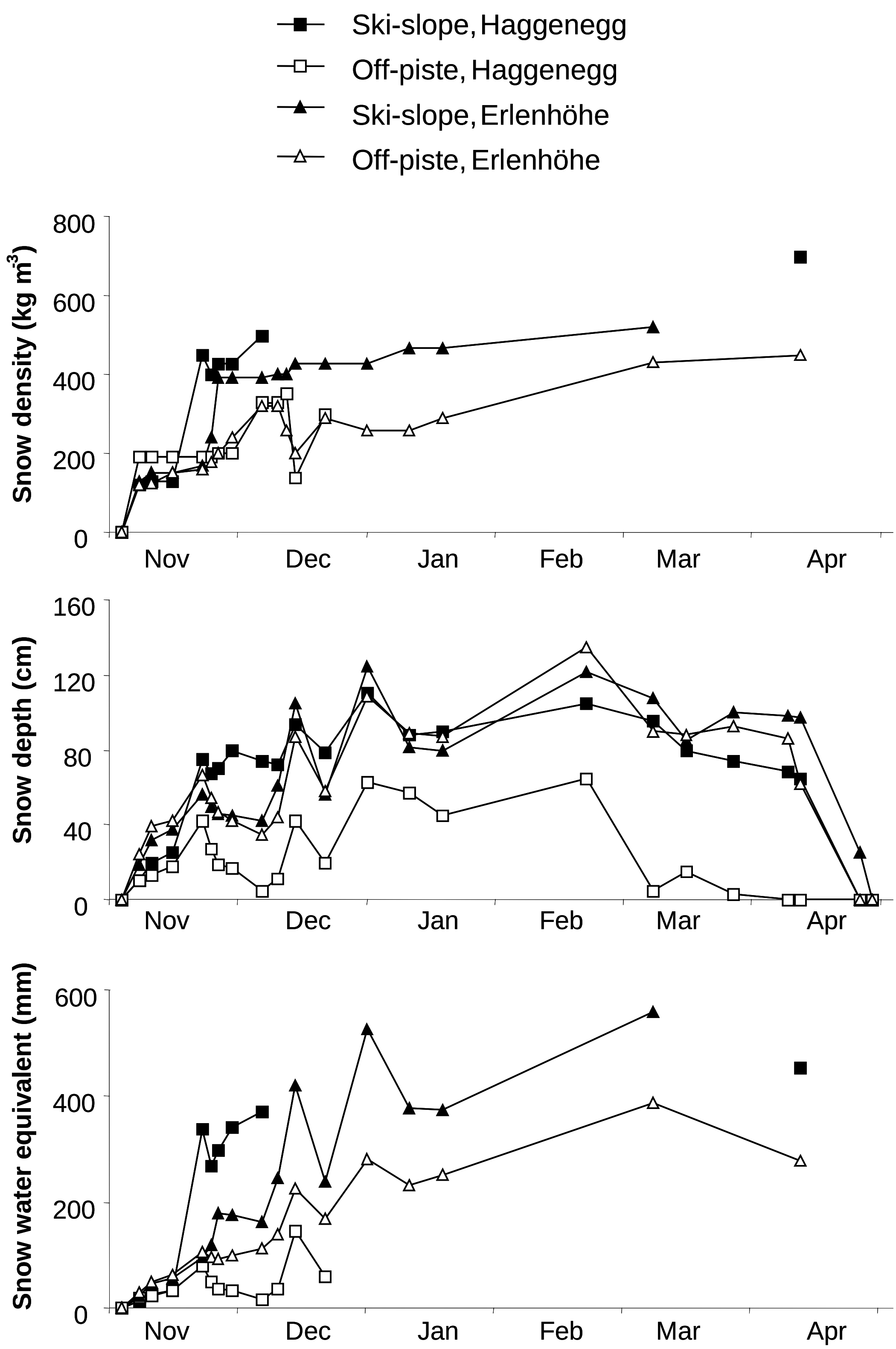

Fig. 1: Measured snow depth (cm), snow density (kg m^{-3}) and snow water equivalent (mm) for the 4 locations during winter 1999/2000.

considerably larger than off-piste due to the additional snow production, whereas at Erlenhöhe both sites had almost the same snow depth. At the off-piste sites the snow disappeared 2 (Erlenhöhe) to 4 (Haggenegg) weeks earlier than on the ski-slopes.

With regard to snow density, we noticed a considerable increase throughout the winter season. The groomed ski-slope at Haggenegg had a mean density of 500 kg m^{-3} in mid-December and gradually compacted to become a mixture of hard snow and pure ice lenses with a maximum density of 700 kg m^{-3}. Meanwhile, the density of the ski-slope at Erlenhöhe never exceeded 500 kg m^{-3}. The mean snow density developed similarly at both off-piste sites. At the end of the season, we measured densities somewhat higher than 400 kg m^{-3}.

5.3.2 Snapshot of the physical snow properties on the ski-slope and off-piste, measured on 12 January

The inspection of the snow properties at the four measurement locations revealed significant differences between ski-slope and off-piste. Thermal conductivity (k_h), density (ρ_{snow}) and hardness (ξ_{snow}) differed by up to one order of magnitude between the investigated locations (Table 3). The man-made snow on the Haggenegg ski-slope showed higher values of k_h, ρ_{snow} and ξ_{snow} than the extensively groomed Erlenhöhe ski-slope. The man-made snow on the ski-slope was on average harder than the one on the ski-slope without artificial snow, and both ski-slope snowpacks were much harder than the off-piste snow cover. Fig. 2 demonstrates the great hardness variability in the profiles of the ski-slopes compared to the natural snowpack. The artificial ski-slope profile from Haggenegg contained repeated layers with a hardness of up to 150 N. This compactness was the result of artificial snow production and intense grooming. The off-piste profile at Haggenegg was the softest and also the most homogeneous profile. Comparing the standard deviation of the repeated hardness profiles at each measuring site we found that the ski-slope with artificial snow had the lowest coefficient of variation (0.4-1.2).

Table 3. Thermal conductivity, snow density and snow hardness measured on 12 January, 2000 (mean values and standard deviation, parenthesised).

	Piste Haggenegg - man-made snow	Piste Erlenhöhe - natural snow	Off-piste Haggenegg	Off-piste Erlenhöhe
Thermal cond. ($Wm^{-1}K^{-1}$)	0.53	0.27	0.14	0.01
	(0.18)	(0.10)	(0.04)	(0.02)
Snow density (kg m^{-3})	530	410	285	200
	(7)	(46)	(64)	(36)
Snow hardness (N)	36.0	11.8	0.6	0.5
	(38.0)	(21.0)	(1.0)	(0.6)

The measurements also confirmed the non-linear relationship between snow hardness, snow density and thermal conductivity. The measured snow densities and the natural logarithm of the snow hardness were correlated ($R^2 = 0.86$, $p < 0.001$). A quadratic function, as used in the

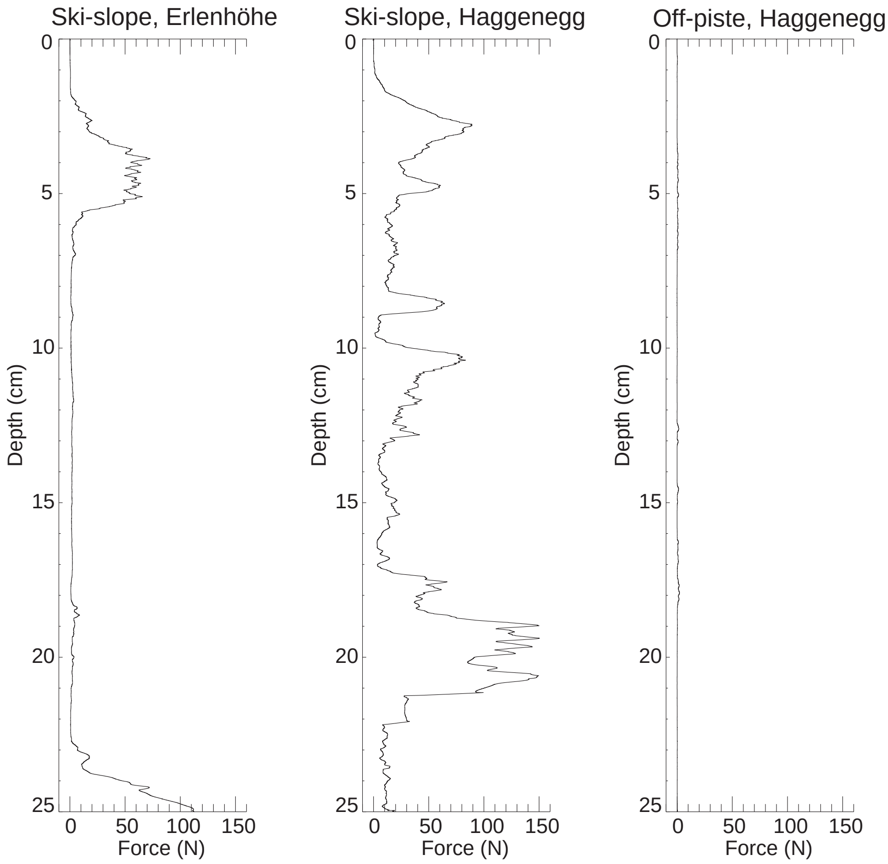

Fig. 2: Examples of penetration force signals measured with the Snow Micro Pen on 12 January, 2000 at three locations of the ski resort Brunni-Haggenegg: (left) ski-slope Erlenhöhe, (centre) ski-slope Haggenegg, and (right) off-piste Haggenegg.

model (Eq. 4), fitted our data set of measured snow densities and thermal conductivity with a R^2 of 0.75 ($p < 0.001$) (Fig. 3) using a s_k -value of 1.8 10^{-6} W m^5 $°C^{-1}$ Kg^{-2}. The correlation of thermal conductivity and snow hardness was weaker than the correlation between the thermal conductivity and the snow density but still highly significant ($R^2 = 0.69$, $p < 0.001$).

5.3.3 Seasonal dynamics of the soil temperatures

In early November when the snow cover started to develop, the soil temperature profiles of the different sites were similar (Fig. 4a). The large temperature gradients between 3 and 20 cm depth indicate that cooling from the overlaying snow was strongly dampened by the large heat capacity of the soil due to its high water saturation (estimated water content: 60-70 %). To the beginning of January, the temperatures in the soil had decreased gradually. No significant differences between ski-slope and off-piste sites could be noticed (Fig. 4b). The coldest temperatures were observed at the Haggenegg off-piste site, which might be explained by the thinnest snow cover or by somewhat drier soil conditions. At the beginning of March, the soil temperatures reached the minima of this winter (Fig. 4c). The soil temperatures at 100 cm depth of both Haggenegg sites were lower than those at the Erlenhöhe sites. As the Alptal is characterised by deep gleyic soils, advective water fluxes of relatively warm ground water or cold melt water could have played an important role to

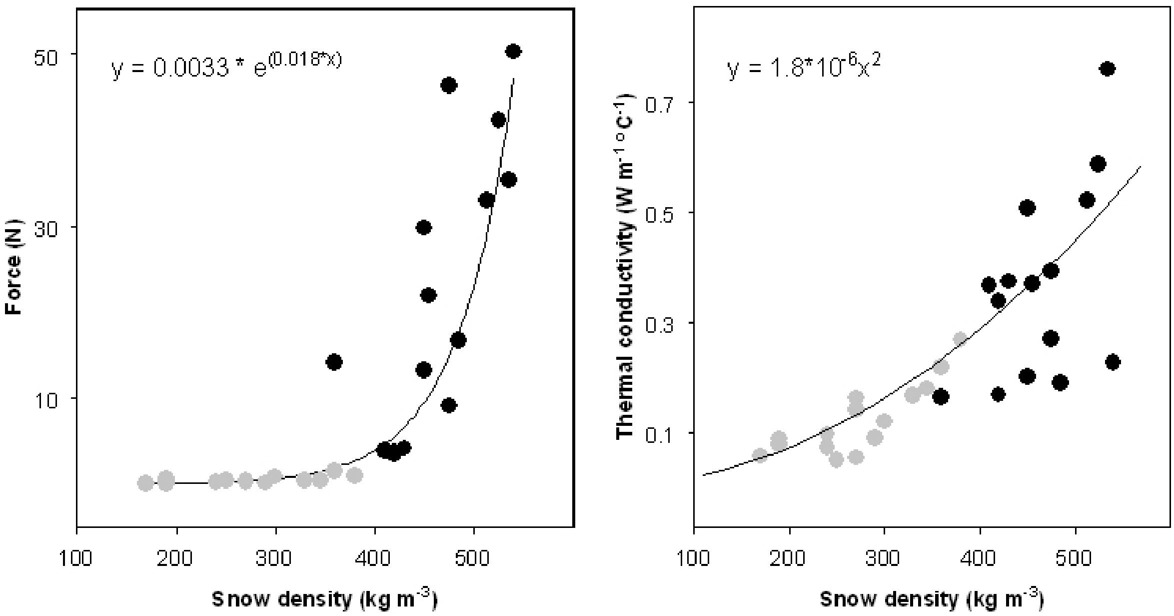

Fig. 3: Relationship between snow density (kg m^{-3}) and snow hardness (N, left), as well as thermal conductivity (W m^{-1} °C, right) measured on 12 January on the ski-slope (dark dots) and in the natural snowpack (light dots).

generate these temperature differences deeper down in the profile. Overall, the soil was never frozen at any of the measured locations throughout the whole season. At the end of the winter season, the snow disappeared on 1 April at the Haggenegg off-piste site followed by a sharp increase of the soil temperature (Fig. 4d). At the ski-slope site Haggenegg a complete melting of the snow cover took four weeks longer time. At Erlenhöhe, the off-piste site was free of snow on 21 April, and the ski-slope two weeks later. As soon as the snow cover had melted completely, the soil surface temperatures increased rapidly. On 1 May, all soil temperature profiles were similar in spite of their different dates of becoming snow-free, except for the off-piste site Erlenhöhe that was still snow covered (Fig. 4d).

5.3.4 Model results

Applications of the numerical model to the two Haggenegg sites are shown in Figs. 5-7. The model produced the seasonal snow depth and snow density development realistically. For the off-piste site (Fig. 5), a satisfactory agreement between simulated and measured snow depth was achieved (R^2 = 0.83). The snow density pattern with layers of low density in early winter and after new snowfalls, which later settled to layers of higher densities, was simulated realistically. A direct comparison of simulated and observed snow density profiles was made for 12 January (Fig. 6). Minimum densities of about 300 kg m^{-3} at the top increasing to maximum densities of about 400 kg m^{-3} at the bottom were simulated in accordance with the measurements. Also for the ski-slope site the model simulated distinct layers of higher or lower densities (Fig. 5) indicating that even regularly groomed ski-slopes are not as homogeneous as one could expect. This heterogeneous snow density profile was confirmed by the measurements of 12 January (Fig. 6). However, the model seems to generally

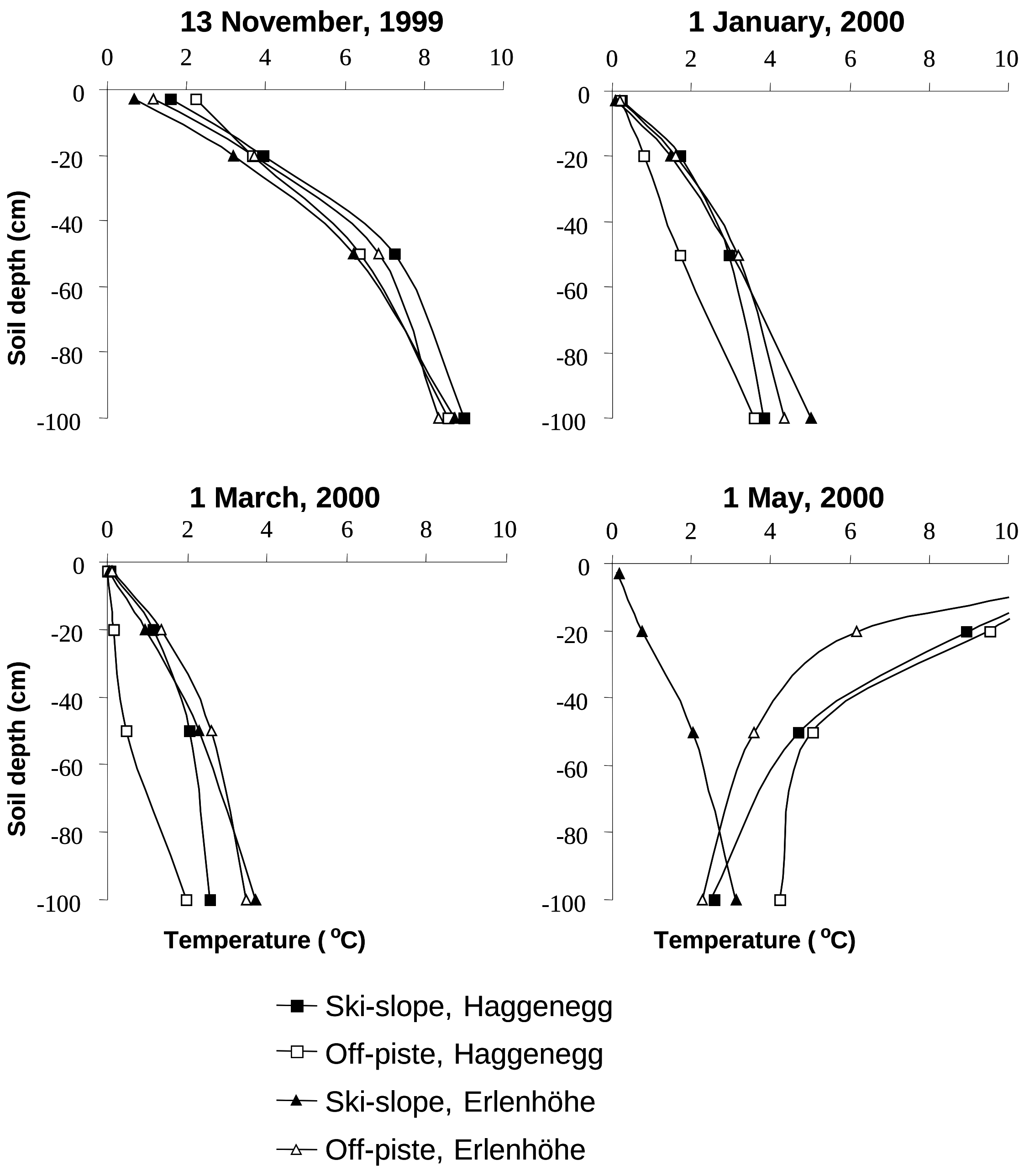

Fig. 4: Measured soil temperature profiles (°C) at four representative dates during the winter 1999/2000.

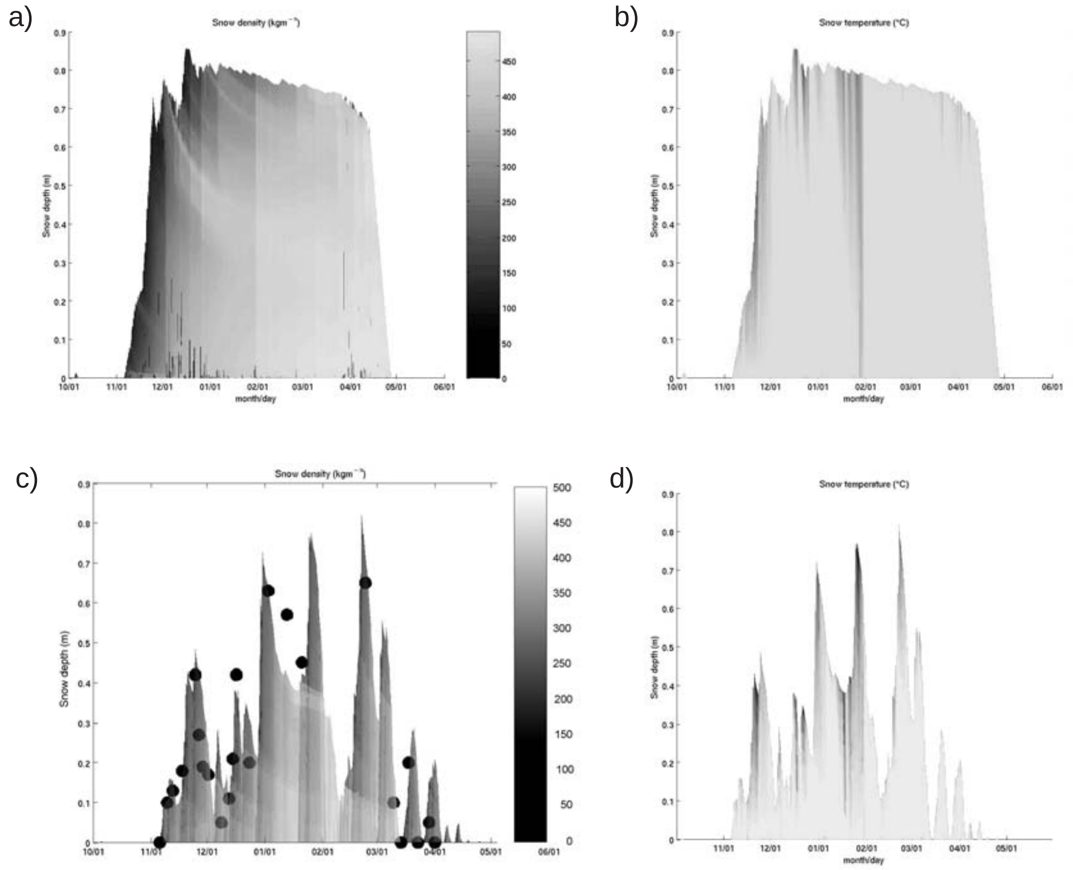

Fig. 5: Simulated snow density profile (left) and snow temperature profile (right) for the ski-slope (top) and the off-piste (bottom) site Haggenegg. For the off-piste site, measured snow depth is indicated with black dots.

underestimate snow density on the ski-slope, especially towards the end of the season Figs. 5 b and d show that for most of the winter season the temperature in the snowpack was close to the freezing point. The model simulated temporal cooling for shorter periods and in the upper part of the snowpack only. The most important freezing event was the second half of January, where the air temperature decreased to less than -15° C. The simulation demonstrates the gradual penetration of the cooling front into the snowpack, however, neither on the ski-slope nor at the off-piste site this cooling front reaches the soil surface for more than one or two days. Consequently, also during this period the soil was not able to freeze. This was also confirmed by the snow temperature sensors at Erlenhöhe that did not register negative temperatures at 3 cm above the soil surface during that period.
The model also reproduced well the dynamics of the soil temperatures. Immediately after the first snow fall in November the temperature decreased rapidly close to the surface and delayed deeper down in the soil (Fig. 7). Compared with the measurements, the temperatures at a depth of 3 cm decreased somewhat too rapidly in the simulation revealing limitations in the representation of the uppermost, organic-rich soil layer. At a depth of 100 cm the cooling of the soil below the snow cover, as well as the re-warming in spring were in satisfying accordance between model and measurements. The model confirmed that the two sites did

not differ significantly with respect to soil temperatures, except at the end of the winter when the snow disappeared one month later on the ski-slope than off-piste.

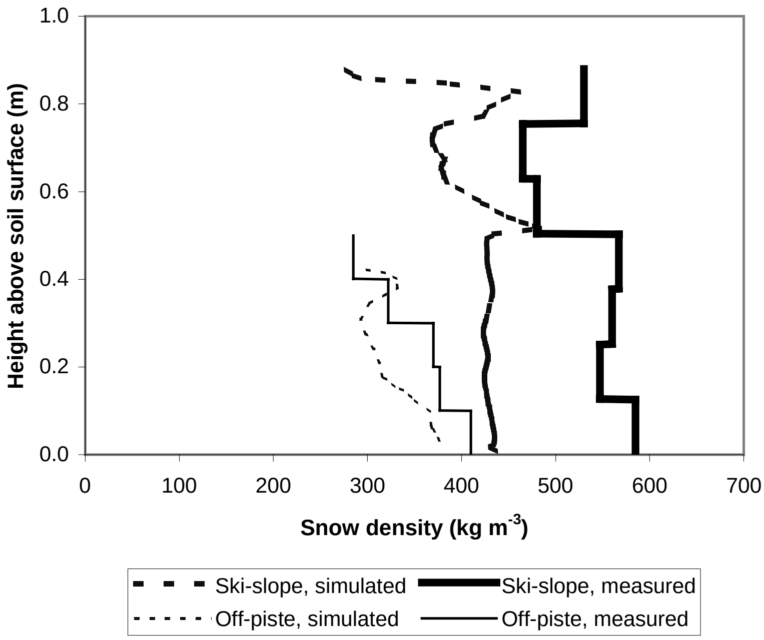

Fig. 6: Simulated and measured profiles of snow density at the Haggenegg ski-slope and off-piste site on 12 January.

5.4 Discussion and conclusions

The conclusion of earlier studies that soils below ski-slopes are subjected to greater cooling than below a natural snowpack is not always appropriate, as the present measurements and simulations have shown. We need to adopt a more differentiated view to this issue. For a relatively low-altitude site such as the Brunni-Haggenegg resort (with a wet soil) it is obviously not decisive how thick, how hard or how dense the snowpack is. Here, the air temperature is normally not cold enough to produce soil frost for more than shorter periods for any kind of snowpack. The only major effect of a ski-slope with additional artificial snow production is the delayed warming at the end of the season. This delay may be 2 up to 4 weeks in a winter such as the one of 1999/2000.

The newly implemented option proved to be a successful strategy for simulating the seasonal snowpack development of a ski-slope. Here, the primary goal is not to simulate the snowpack itself but to provide best possible boundary conditions to the soil when the task is to calculate heat and water fluxes below a ski-slope snowpack. By adjusting the simulated snow depth to real measurements we may on the one hand eliminate the errors related to the precipitation measurements which often may be considerable, especially for snow precipitation in an environment with strong wind influence, and on the other hand we may include artificial manipulations that can not be measured or for which there is no quantitative information available. Lehning and others (2002) demonstrated that this model strategy was also successfully applied in high alpine areas where automatic weather stations provide the input data (snow depth) with no manual measuring throughout the whole winter season.

The main uncertainty related to this simulation procedure is how to handle the snow density at the moment when snow is added or removed. Either we could assume that the additional snow is new-fallen snow with a corresponding low density, or we believe that the added snow amount corresponds to an artificial snow production followed by a mechanical grooming, which would mean that the snow density should be in the same order as before the snow addition. In the simulations presented we proceeded according to this second case, and it showed that this was an appropriate way to describe the snow density on the ski-slope.

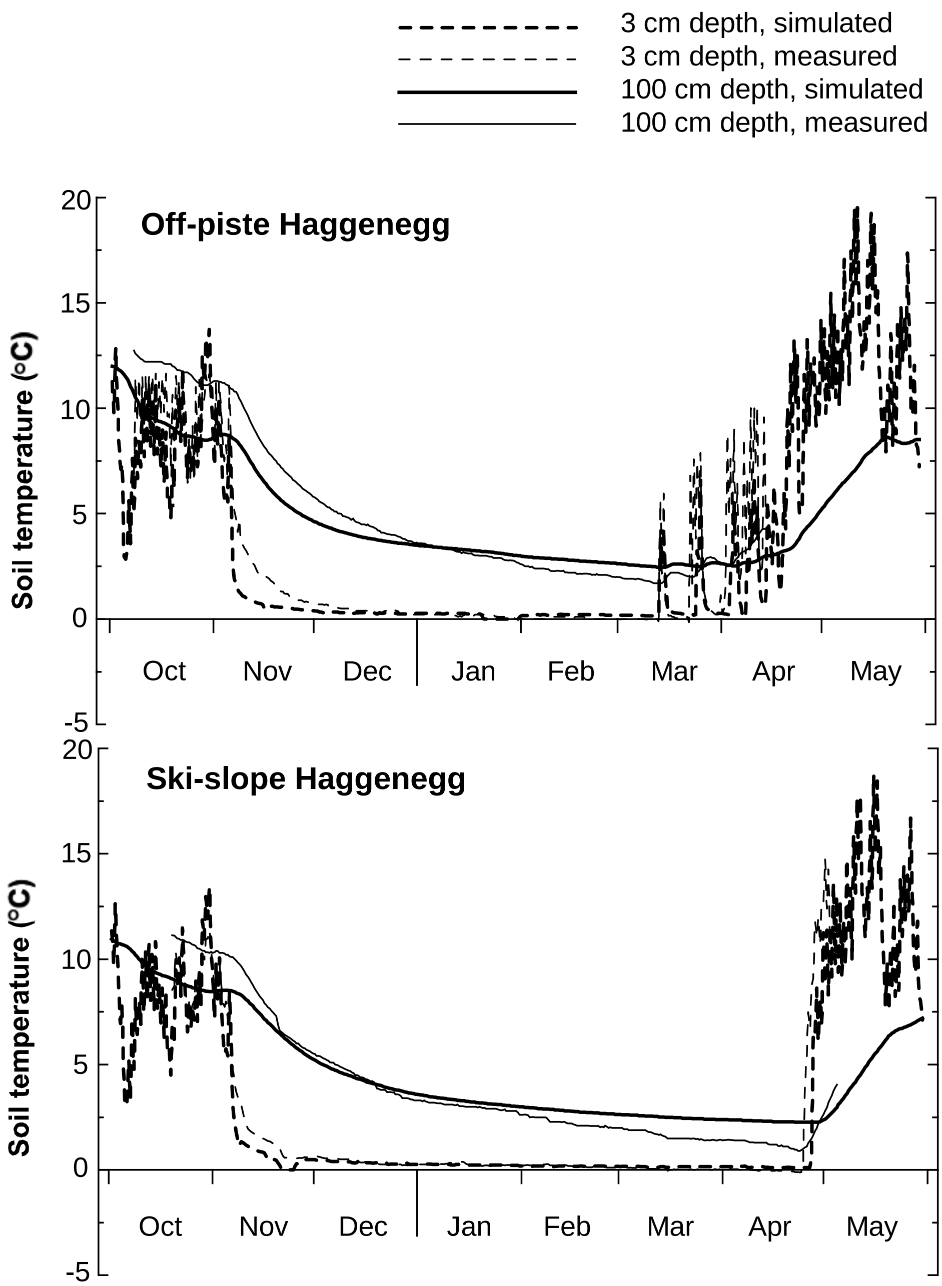

Fig. 7: Simulated and measured soil temperatures at a depth of 3 and 50 cm for the ski-slope and the off-piste site Haggenegg.

On the other hand, Lehning and others (2002) assume in their model that all added snow is natural snowfall. They calculate the density of new-fallen snow with a multiple linear regression model including air temperature, surface temperature, relative humidity and wind speed that was fitted to an extensive data set from the Swiss alps (Davos, 1500-2500 m a.s.l.). Having a suitable tool for calculating the impact of the snow cover on the soil temperatures, such as the COUP-model presented here, is of course only the first step towards a comprehensive assessment of environmental impacts from intensively prepared ski-slopes. Our study has shown that there is no unique answer as to whether modified snowpacks alter the thermal conditions in the soil or not, and consequently there will be no unique answer as to how much ski-slopes influence the composition and magnitude of plants. For ski resorts where new snow-gun installations are planned it will be necessary to make individual calculations and estimations as to what extent artificial snow production and snow compaction may alter the soil ecology. For this purpose further efforts have to be taken to incorporate plant growth response to thermal conditions, as well as soil chemical aspects into the model to allow site-specific conclusions to be drawn.

5.5 Acknowledgement

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Chapter 6

Snow stratigraphy measured by snow hardness and compared to surface section images

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Abstract

Snow hardness is one of the most important parameters in the study of snow. Hand, ram and micro penetrometer measurements were taken in seven snow pits. The snow profiles are analysed in terms of the three hardness tests and surface section images are used as an objective reference. From the images we established the stratigraphy in terms of layers and layer boundaries at high spatial resolution and compared the stratigraphy to the hardness profiles. The hand test captured 80 % of all layers and layer boundaries and the ram test 60 %. The micro penetrometer captured the stratigraphy more complete than hand and ram profiles. The hand and ram profiles are practical but coarse methods to record the snowpack stratigraphy and snow hardness. Important details are missed, for example thin hard and soft layers which are highly relevant to avalanche formation. Differences in soft snow are resolved by the micro penetrometer, which is problematic or impossible with hand and ram tests. The hand and ram hardness tests indicate systematically greater snow hardness than the micro penetrometer. The surface sections and micro penetrometer profile show a much more stratified snowpack than revealed in the classical snow profiles. Quantitative evaluation of mechanical and textural snowpack properties requires methods that have a spatial resolution of at least 1 mm. Since the main heat and mass fluxes are perpendicular to the snow surface, the much stronger stratification now revealed has a large impact on vapour transport. Avalanche formation (metamorphism in thin layers), hydrology (water flow) and electro-magnetic models (microwave emission and radar) are examples where a highly resolved snowpack stratigraphy will be important.

6.1 Introduction

The snowpack consists of numerous layers and layer boundaries that are related to the different snowfall, melting, wind erosion and deposition periods as well as to the metamorphic processes within the snow cover. A layer is a stratum of snow that is different in at least one respect from the strata above and below (Colbeck et al., 1990). A layer boundary is an interface or a transition between two adjacent layers (Colbeck, 1991). Snow stratigraphic profiles represent the varying properties of layers and layer boundaries in a snowpack. Layers are therefore the representative elementary volume (Bear, 1972), assumed to have “infinite” horizontal extension and homogeneous physical and mechanical properties.

Depending on the aim of the snowpack investigation, the classification is focused on different properties. A classification of snow grains yields a representation of the metamorphic state of the snowpack. Sturm et al. (1995) developed a snowpack classification system according to climatic parameters and physical snow parameters. Snowpack stratigraphy and snow properties play also an important role in questions of electromagnetic properties (Mätzler, 2000, Wiesmann et al., 2000, Boyarskii and Tikhonov, 2000). Permeability is an important variable for water and chemical fluxes in the snowpack and it is strongly dependent on the stratigraphy and on the snow properties (Albert et al., 2000, Albert and Perron, 2000). The transfer of mass and energy through a snowpack strongly depends on the number and the properties of the layers in the snow. Most fluxes, such as heat, air and mass are perpendicular to the snow surface. Kirchhoff’s law, in analogy to direct current circuits, explains the significance of thin horizontal layers. The total resistance in a circuit with series-connection is the sum of all single resistances whereas the reciprocal of the total resistance of a circuit with parallel connection is the sum of the reciprocals of the single resistances. It follows that a resistor (analogous to a layer) has a much greater effect when series-connected (analogue to fluxes perpendicular to the layering) than when parallel connected (analogous to horizontal fluxes). While the total heat flow can be averaged, processes like vapour flow are a very localised process. Because they change the texture by recrystallisation, numerical models not representing a small scale layering will be misleading.

To measure the mechanical hardness of a snowpack, Haefeli (Bader et al., 1939) developed the Swiss ramsonde from penetrometers used in soil mechanics. This was the first method to gain objective mechanical data from the snowpack. Due to its relatively large tip (4 cm diameter) the ramsonde is unable to detect thin and soft layers often responsible for the formation of avalanches. Abele (1963) showed the relationship of ram hardness and unconfined compressive strength in hard, processed snow and compressive strength is related to snow density. Hardness profiles were characterised for stability evaluation by de Quervain and Meister (1987) and this classification was extended by Stoffel et al. (1998) and Schweizer and Lütschg (2001).

The International Classification for Seasonal Snow on the Ground (Colbeck et al., 1990) is the standard in describing the most important features of seasonal snow covers. It is the basis for the snowpack interpretation methods used by snow researchers and operational avalanche warning services. A snow profile is a one-dimensional observation of the stratigraphy that is determined by a hand hardness test and visual inspection. Snow hardness, grain morphology and grain size are measured in the defined layers. Hardness can vary greatly between observers because it is subjective and there is no absolute hardness reference. Geldsetzer and Jamieson (2001) showed the relationship of hand hardness and density. Colbeck (1991) emphasizes that is is important to investigate the spatial variability of the snowpack properties and notes the lack of methods. The physical properties of the layers and the stratigraphy play an important role for questions of weak layer development and non-uniform infiltration of melt water (Colbeck, 1991). To understand thermal fluxes within the snowpack, Sturm and Johnson (1991) as well as Arons and Colbeck (1995) point to the importance of snow micro structure and snowpack variability. Investigations of snowpacks at forested sites (Imbeck, 1987, Albert and Hardy, 1993) revealed complex systems of layers

whose properties are difficult to capture with the classical methods. Shapiro et al. (1997) state that it is necessary to measure and interpret the snow micro properties of snow under loading to understand the processes of avalanche release. The term micro scale is used here according to the definition of Arons and Colbeck (1995). It describes the level of the diameter of a grain or of parts of a grain and we use up to 1 mm as maximum size at the micro scale. The meso scale is from mm to cm, and the macro scale is the scale of the thickness of a layer to the thickness of the entire snowpack (Arons and Colbeck, 1995).
Because the stratigraphy is central to all processes acting within the snowpack and thin and soft layers are highly relevant we propose that it is necessary to capture the snow stratigraphy based on highly resolved and objective mechanical hardness measurements that entail additional information on the snow texture. By measuring snow hardness with the SnowMicroPen (Schneebeli and Johnson, 1998) these requirements can be fulfilled (Johnson and Schneebeli, 1999, Schneebeli et al., 1999). The SnowMicroPen is applied for the first time along with the classical methods in dry, natural snowpacks. Hand hardness, ram hardness and micro penetrometer hardness profiles are compared. We show the fundamental differences of the three methods. The stratigraphy is compared and the hardness profiles are correlated. The suitability of the methods to capture the snowpack properties and the hardness and texture variations in a snow profile is discussed.

6.2 Methods

6.2.1 Snow hardness

Snow hardness is the resistance to penetration that has the dimension of a force. It is given by the penetration of a material that is harder than snow. We are comparing three hardness measurement methods for snow: hand hardness, ram hardness and micro penetrometer hardness. Tab. 1 contains the resolution and operating characteristics of each method.

Hand hardness: The hand hardness test is a somewhat subjective, observer dependent manual penetration test that is used to establish the snowpack stratigraphy in a snow pit. The observer measures hand hardness by pushing the fist or 4 fingers or 1 finger or pencil or knife with a given force (about 50 N) and parallel to the layer into the snow and by sensing the resistance by hand (Colbeck et al., 1990). The vertical resolution of the hand hardness test is not constant. It depends on the elementary area of the measuring device and ranges from 0.3 to 7 cm, where knife blade is 0.3 cm, pencil is 1 cm, 4 fingers and 1 finger is 2 cm and fist is 7 cm. The values of the hand hardness are on an ordinal scale. Captured are primarily the hardness differences of the layers in the one profile under investigation. Hardness itself can vary greatly between observers since the pushing force and the elementary area are subjective. There is no absolute hardness reference. The hand hardness levels were correlated to ram hardness by deQuervain (1950) and this correlation was modified in the International Classification for Seasonal Snow on the Ground (Colbeck et al., 1990). The latter correlation is used for the conversion of hand hardness into ram hardness, shown in Tab. 2.
Ram hardness: The Swiss ramsonde (Bader et al., 1939) is the classical instrument to objectively measure snow hardness. The ramsonde is driven into the snow by mechanical hammer blows on top of the probe. The vertical spatial resolution of the ram test is limited by the size of the measuring cone (4 cm diameter) and is at best 1 cm. It is well known but was

Table 1: Characteristics of hand, ram and micro penetrometer hardness methods.

	Vertical spatial resolution: [cm]	**Hardness resolution [N]**	**Reference area: "elementary area" of measurement: [mm^2]**	**Deformation velocity [m s^{-1}]**	**Pene-tration direction**
Hand Hardness	0.3 to 7	10^1	variable 1 to 5000	variable 0.01 to 0.1	horizontal
Ram Hardness	1.5 to 5	10^1 to 10^2	constant 1250	variable 10^{-4} to 1	vertical
SnowMicroPen Hardness	0.1	10^4	constant 20	constant 0.02	vertical

Table 2: Hand hardness indices and corresponding ram hardness (Colbeck et al, 1990, extended).

Hand test	**fist**		**4 fingers**		**1 finger**		**pencil**		**knife**	
Class (Colbeck et al., 1990)	very low		low		medium		high		very high	Ice
Ram classes [N] (Colbeck et al., 1990)	0-20		20-150		150-500		500-1000		> 1000	
Hand index (SLF)	1	1.5	2	2.5	3	3.5	4	4.5	5	6
Ram values [N] (this study)	3	20	65	150	280	500	725	1000	1500	2000

never published that there is a positional lag in the vertical distances in the ram profile. The hardness resolution is limited by the weight of the probe itself and the weight of the hammer. Gubler (1975) showed the dependency of the ram hardness on the speed of deformation, which is not constant during the ram hardness measurement. Gubler (1975) also pointed to the problem of the considerable energy loss at the connections of the individual parts of the ramsonde and developed a new ram hardness equation that partially accounts for these energy losses.

Micro penetrometer hardness: The SnowMicroPen measures the snow hardness and snow texture at the millimeter scale (Schneebeli and Johnson, 1998, Schneebeli et al., 1999). The instrument records the penetration resistance on a small tip with high spatial resolution (4 µm) and high force resolution (0.005 N). The fundamental idea is that a quasi-continuous recording, small diameter penetrometer will connect more directly to the snow micro properties than large diameter cone penetrometers do (Johnson and Schneebeli, 1999). To show the reproducibility of SnowMicroPen field measurements, micro penetrometer hardness profiles were taken at a distance of 0.1 m. The force signal was smoothed with a moving average filter calculated for each millimeter. Three micro penetrometer profiles and the difference between the central measurement (A) and the two neighbouring, 0.1 m distant measurements (B and C) are shown in Fig. 1. The largest differences occur at very hard layers, such as the crust at about 1 m profile depth. These differences are attributed to the small scale spatial variability. The heterogeneity of a crust's boundaries and shape is shown in the surface image in Fig. 2.

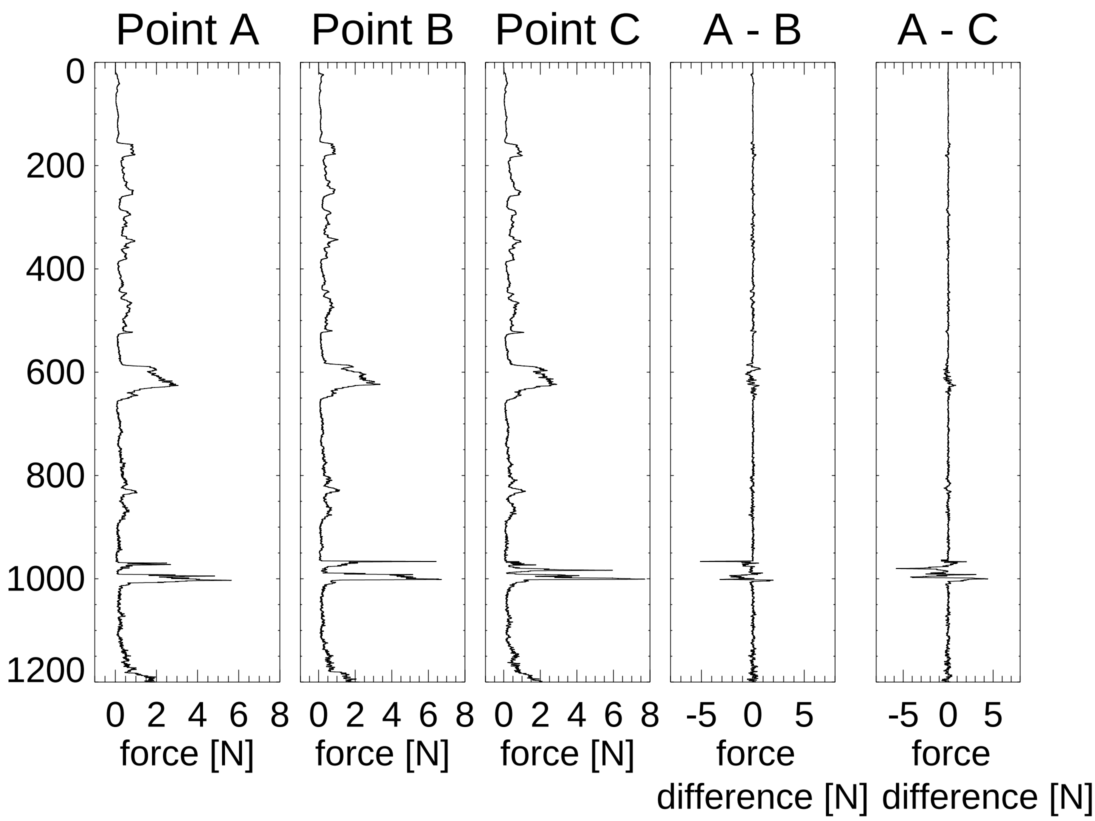

Fig. 1: Three micro penetrometer hardness profiles from the study plot Weissfluhjoch, Davos, point A, B and C with a spacing of 10 cm show the reproducibility of SMP measurements. Differences, A-B and A-C, occur at hard layers and crusts and are attributed to small-scale spatial variability of the snow properties in these layers.

6.2.2 Snow samples and snow micro structure images

To analyse the undisturbed snow stratigraphy and snow micro structure, 18 snow samples (7 cm x 7 cm x 5 cm) were taken from sub-areas of the four profiles taken during the winter 2001/02. To conserve the fragile snow structures during transportation and laboratory processing, the snow samples were filled with dimethyl phthalate and frozen. The surface section method described by Good (1987) was improved to yield better contrast and higher effective resolution (10 µm) in the digital images. Still higher resolution is obtained by producing image-mosaics, where a sample surface is photographed in five sections, which are later reassembled digitally. The images document the actual stratigraphy and the grain properties of the 7 cm high and 1 cm wide sub-area of the sample. The location of the samples within the snow profile is marked in Fig. 3 with horizontal black bars above the x-axis.

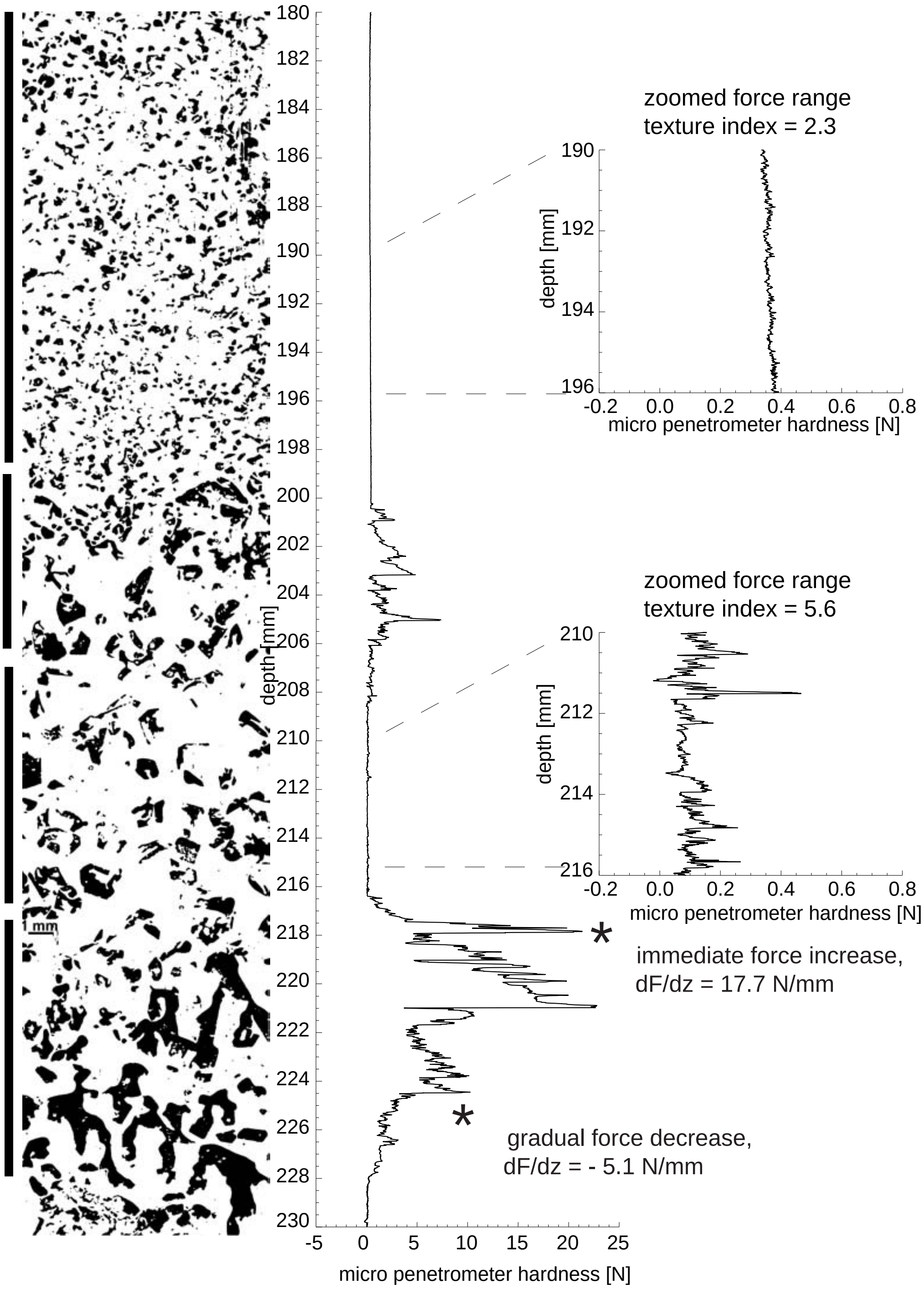

Fig. 2: Snow profile from 9 Jan. 2002: serial sample containing two crusts (black is ice). The black vertical bars on the left mark the layers as they are depicted from the surface image. Their interruptions mark the layer boundaries. This is an example where the stratigraphy established by hand hardness agrees with the surface image. Snow texture and hardness gradients at layer boundaries are calculated from the micro penetrometer hardness signal.

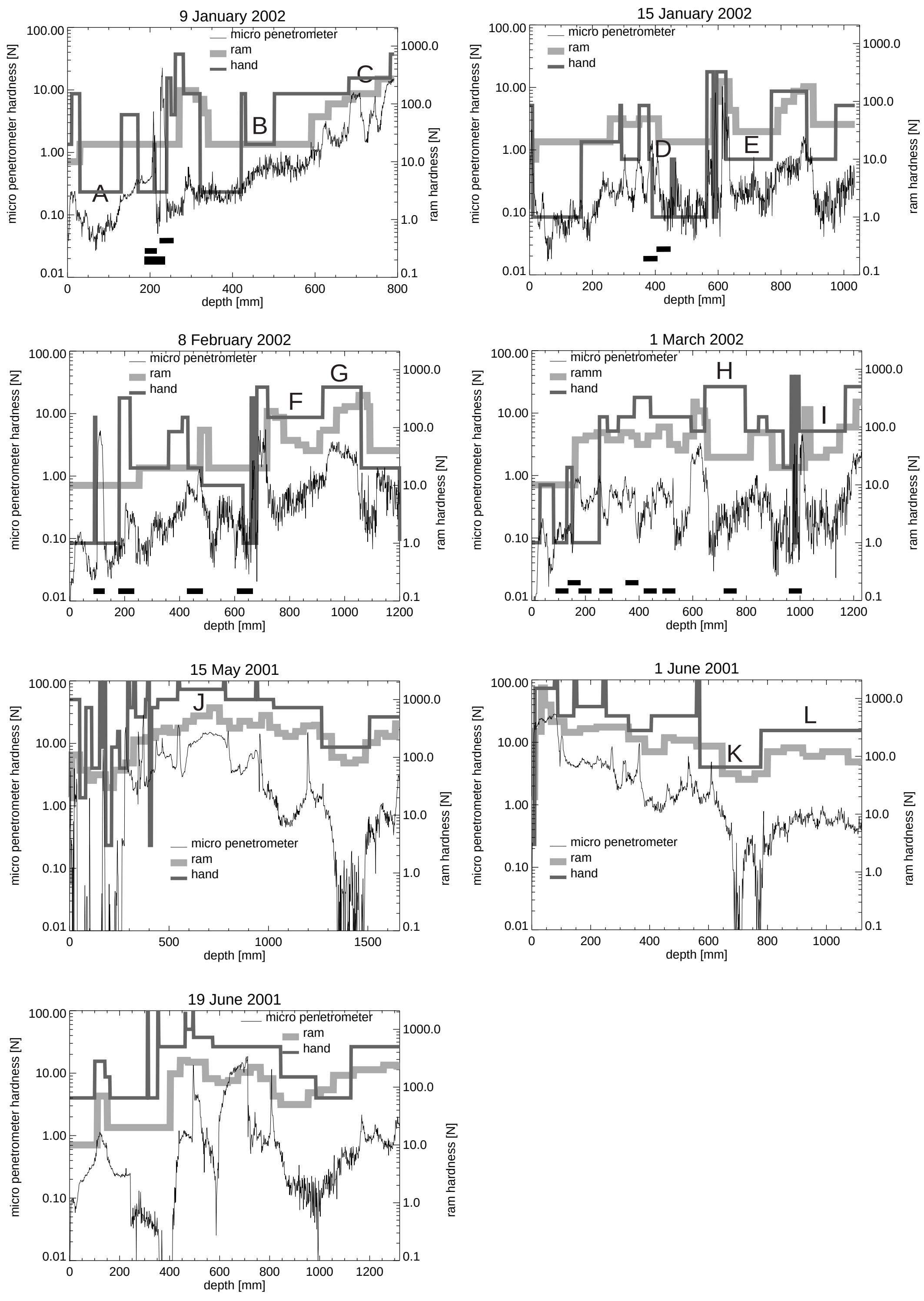

Fig. 3: Snow hardness profiles measured by hand hardness and ramsonde (right y-axis) and by micro penetrometer (left y-axis, smoothed with moving average over 1 mm window). The logarithmic y-axes are chosen to get a better resolution in the lower hardness values. Differing in about one order of magnitude, the three hardness profiles are drawn on the same graph and can be approximately compared at these scales. The horizontal black bars above the x-axes mark the positions where the snow samples were taken. The letters A to L mark layers discussed in the text.

6.2.3 Comparison of methods

The comparison of the hardness profiles is based on the stratigraphy reconstructed from the highly resolved surface images of the snow samples. The surface images serve as an enlargement of the snow profile sub-areas. The position and number of layers and layer boundaries are determined from the surface images by visual inspection of changes in snow density and grain properties. Fig. 4 shows an example of this classification on the left of the surface image where the black vertical bars indicate the layers and the interruptions of the bars indicate the layer boundaries. The presence or absence of these features in the hardness profiles is determined and the occurrences are counted and compared. Hardness differences at layer boundaries that are present amongst the three hardness profiles are correlated. The layer thickness is analysed and discussed. Selected stratigraphic layers, as they are defined by the hand hardness method are compared to the micro penetrometer profile in detail. Snow hardness is analysed by the comparison of the cumulative probability distributions and general statistical distributions of the hardness values of the three methods.

6.3 Data

Hand hardness, ram hardness and micro penetrometer hardness profiles were taken adjacent to one another in seven snow pits during the winters of 2000/01 and 2001/02. The measurements were done in winter (dry) and spring (moist/wet) snowpacks. Two slope profiles were taken in the area of Choerbschhorn, nearby Davos, Switzerland on January 9 and 15, 2002. Two flat field profiles were taken on February 8 and March 1, 2002 at the study plot Weissfluhjoch, Davos. The much harder spring profiles were measured during the winter 2000/01 on May 15, June 1 and June 19, 2001 at the study plot Weissfluhjoch, Davos. From the four profiles of the winter 2001/02, 18 casted samples for serial section analysis were taken from sub-areas of the profile and transferred to the cold laboratory for preparation and analysis. A summary of the field experiments is given in Tab. 3.
In the seven profiles, the vertical distribution of the hardness of snow, as measured by the hand and ram tests and the SMP, differed strongly (Fig. 3). The hardness values are plotted on logarithmic y-axes. Differing one order of magnitude, the hardness values are shown on the same graph and can be compared at these scales. The logarithmic y-axes are chosen to get a better resolution at lower hardness. To verify the reproducibility of micro penetrometer measurements, we took additional profiles on March 1, 2002 at the study plot Weissfluhjoch, Davos. These measurements were taken from a precision measuring table, which is mounted on skis, in a regular 0.1 m grid. This measuring table assures exact horizontal spacing and undisturbed snow conditions for the closely spaced measurements. The complete dataset, penetrometer files, image files and snow profiles are available at http://www.slf.ch/research/snowstratigraphy/data-de.html.

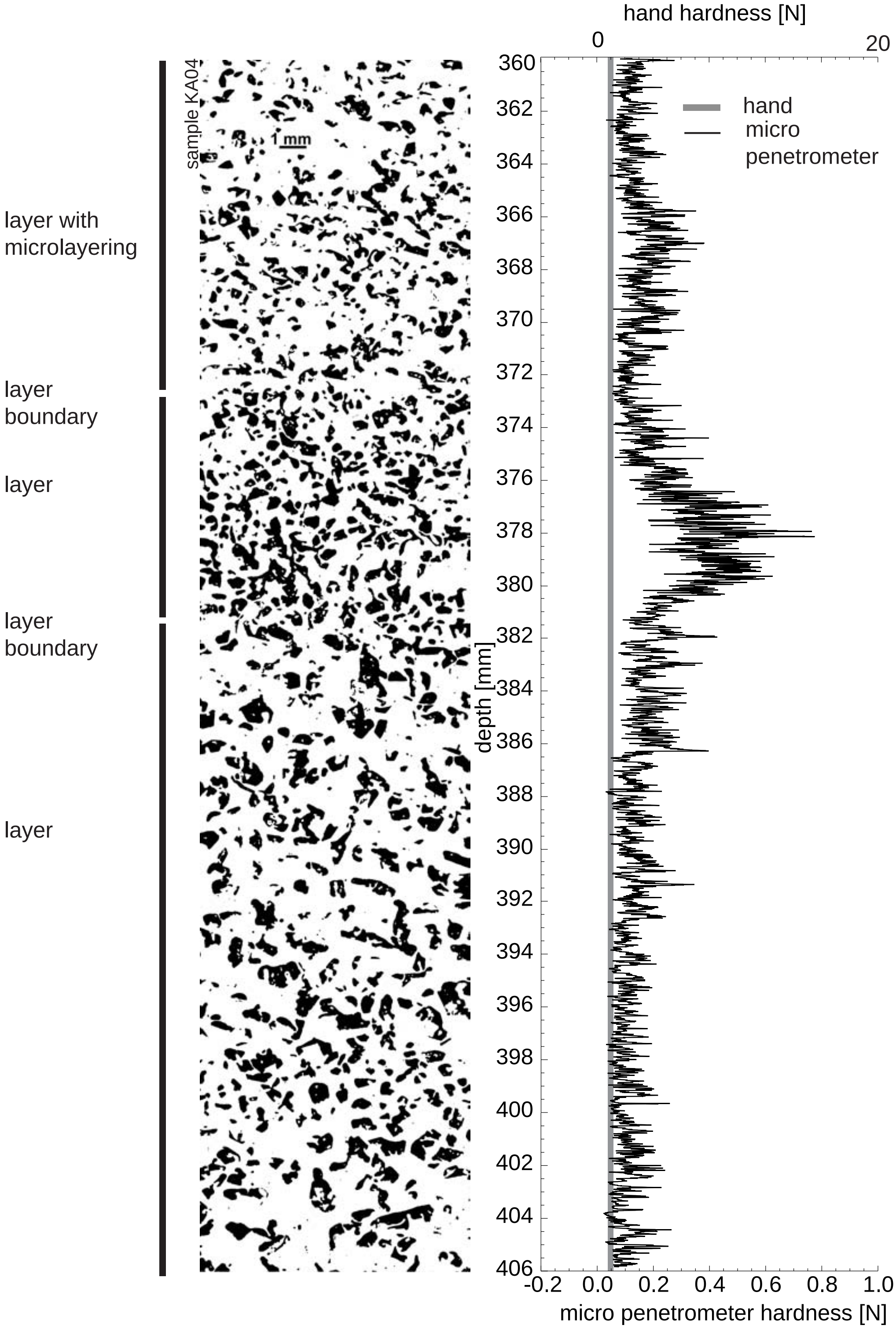

Fig. 4: Snow profile from 15 Jan. 2002: surface image containing thin, harder layer and micro layering (black is ice) and the corresponding hand and micro penetrometer hardness. This is an example where the stratigraphy established by hand hardness disagrees with the surface image because the thin, harder layer is missing.

Table 3: Field experiments where hand, ram and micro penetrometer hardness profiles and snow samples were taken.

Date	*Location*	**Casted snow samples**	**Snow sample location in snow profile (mm from snow surface)**
09.01.02	Choerbschhorn - slope	KA01	180-230 mm
		KA02	180-210 mm
		KA03	210-240 mm
15.01.02	Steintälli - slope	KA04	350-410 mm
		KA06	410-460 mm
08.02.02	Weissfluhjoch study plot - flat	vf01_1	90-120 mm
		vf01_2	180-230 mm
		vf01_3	430-480 mm
		vf01_4	610-660 mm
01.03.02	Weissfluhjoch study plot - flat	vf02_1	80-130 mm
		vf02_2	130-180 mm
		vf02_3	170-220 mm
		vf02_4	250-300 mm
		vf02_5	350-400 mm
		vf02_9	420-470 mm
		vf02_6	490-540 mm
		vf02_7	720-770 mm
		vf02_8	970-1015 mm
15.01.01	Weissfluhjoch study plot - flat	none	
01.06.01	Weissfluhjoch study plot - flat	none	
19.06.01	Weissfluhjoch study plot - flat	none	

6.4 Results and discussion

The seven profiles are shown in terms of their hand, ram and micro penetrometer hardness profiles in Fig. 3. The hardness profiles from each method are very different in such that ram hardness is more variable than hand hardness and micro penetrometer hardness is even more variable than the other two. The hand profile has a better spatial resolution than the ram profile and it contains more extreme values than the ram profile. Hand hardness is not objective because the observer focuses subjectively on interesting layers, which has proven practical for the observation of potential fracture layers. The differences in the hardness profiles also stem from the underlying, different deformation processes inherent in each method. Hand hardness is measured with devices with variable reference areas and shapes but at roughly constant deformation velocity. This results in different deformation processes even amongst the hand hardness classes. Ram hardness is measured with a constant reference area but at variable deformation velocities. Because the SnowMicroPen has both a constant reference area and a constant deformation velocity, and it also has the highest spatial and force resolution it is considered the most reliable hardness profile. The results of the following comparisons of layers and layer boundaries in the hardness profiles as well as the comparisons of the statistical distributions of the hardness values support this proposition.

6.4.1 Representation of layers and layer boundaries

In the surface images of the 18 snow samples we determined 41 snow layers and 17 layer boundaries. We compare this count to the occurrences in the different hardness profiles. Of all 41 layers, 25 (61 %) were recorded in the ram hardness profiles, 31 (76 %) in the hand hardness profiles and all were recorded in the micro penetrometer profile. Of all 17 layer boundaries, 9 (53 %) were recorded in the ram hardness profiles, 14 (82 %) were recorded in the hand hardness profiles. From the micro penetrometer hardness profile all layers and layer boundaries could be identified. In the case of three layer boundaries, the sign of the hardness difference was opposite in micro penetrometer and hand hardness profiles. A possible explanation is a recording error in the manual profile. Of all 17 determined layer boundaries, 14 coincide in the hand and micro penetrometer profiles, 9 coincide in the micro penetrometer and ram profiles, and only 5 coincide in the hand and ram profiles. An example where the stratigraphy established by hand hardness agrees with the surface images is shown in Fig. 2. An example where it disagrees is shown in Fig. 4 where within a soft layer a thin, harder layer exists that does not appear in the stratigraphy. The micro penetrometer profile entails a layer with a hardness difference of 0.4 N.

The correlations of hardness differences at layer boundaries between hand and micro penetrometer hardness, between ram and micro penetrometer hardness and between ram and hand hardness are shown in Fig. 5. The correlation (R^2 0.34, p-value 0.01) between the hardness differences in hand and micro penetrometer profile is statistically significant. The ram hardness measurement method produces significantly different hardness discontinuities at layer boundaries than hand and micro penetrometer. The layer boundaries in the hand hardness profiles are predefined discrete steps of the hardness classes. The resolution of the ram hardness is dependent on the operator and on the snow hardness itself. Generally, the ram hardness resolution is low in soft snow. Gradually changing properties at layer boundaries are impossible to capture by hand hardness and difficult by ram hardness. Sharp hardness discontinuities in a snowpack are one of the controlling factors in avalanche formation (Schweizer and Lütschg, 2001). The SnowMicroPen has enough resolution to account for the gradient of a hardness discontinuity at the millimeter scale. For two exemplary layer boundaries, on top and below a crust, this is shown in Fig. 2. The gradient of the force increase at the upper crust boundary and decrease at the lower crust boundary are calculated. From a linear fit to the micro penetrometer force data over a distance of 1 mm (250 data points) the average force gradient is calculated. The average force gradient in the millimeter above the crust to the first peak of the crust is 17.7 N mm^{-1} and for the discontinuity from the last peak of the crust to the millimeter below the average force gradient is -5.1 N mm^{-1}. With this method the magnitude of the hardness differences at layer boundaries can be quantified from micro penetrometer profiles. The absolute force record and the snow texture preceding or following a hardness discontinuity are also important factors when evaluating hardness discontinuities. In Fig. 2, two exemplary profile sections are zoomed and the texture index is calculated for each section. The texture change can be observed in the surface images and is reproduced in the calculated texture index. The higher texture index is an indication for lower textural stability (Schneebeli et al., 1999).

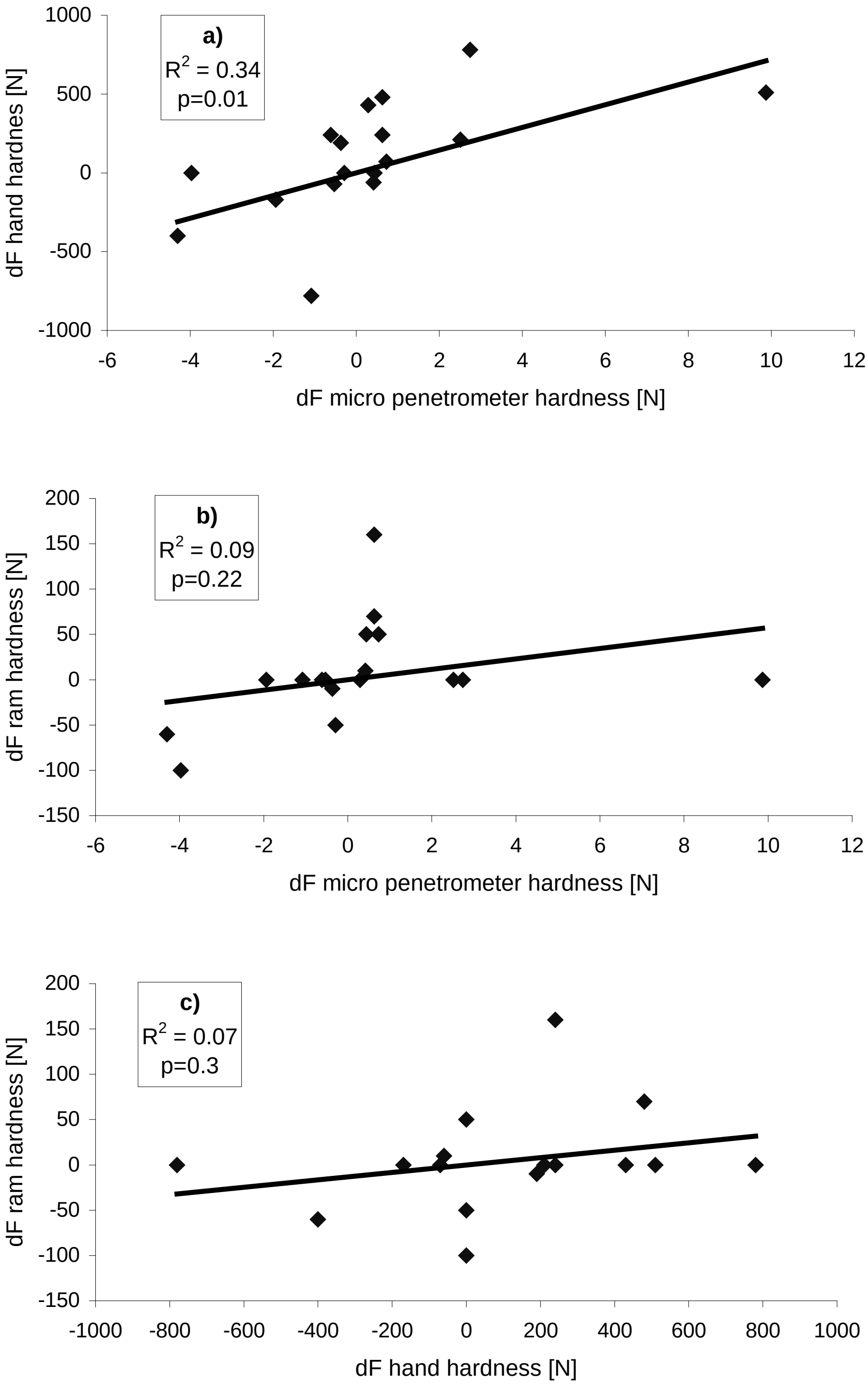

Fig. 5: The hardness differences (dF) at layer boundaries between: a) hand and micro penetrometer hardness are correlated ($p=0.01$), b) ram and micro penetrometer hardness and c) ram and hand hardness are uncorrelated (p-values at 95 % confidence level).

6.4.2 Layer thickness

The relationship of hand and ram hardness levels and layer thickness is given in Fig. 6. The ram hardness profile is biased by thin hard and thick soft layers. The hand hardness profiles show no bias in layer thickness. The relative frequency of layer thickness in all 7 ram and hand profiles is shown in Fig. 7. The thinnest recorded hand layer is 0.1 cm and the thickest is 37 cm. The thinnest recorded ram layer is 1 cm and the thickest is 25 cm. 67 % of all hand layers and 92 % of all ram layers are up to 7 cm thick. The hand layers up to 7 cm thickness are shown in terms of the required elementary thickness (r_thick) for a measurement of the hand hardness level in the inset of Fig. 7. 50 % of the thinnest, up to 1 cm thick layers are estimated because the reference thickness of the device is too large. 30 % of up to 1 cm thick layers are recorded as ice lamellas. Considering all the hand layers with a thickness from 0 to 7 cm (except for the ice lamellas), we find that 35 % of these layers do not have the required thickness for the hand hardness level and therefore have hand hardness estimations. Thin layers occur frequently in all hand and ram profiles. With the hand hardness test, which is the basis for the snowpack stratigraphy, it is not possible to accurately determine the hand hardness of 25 % of all recorded hand layers.

6.4.3 Homogeneity of layers compared to micro penetrometer hardness

A selection of stratigraphically homogenous layers is exemplarily compared to the micro penetrometer hardness profile. These layers are marked in Fig. 3 by the capital letters A to L above the hand hardness layer. Layer A varies between 0.03 to 0.11 N around a mean value of 0.06 N. Here, the signal is superimposed by a higher frequency signal with a wavelength of a few millimeters, with similar amplitude as the main increase. This pattern is typical for wind influence during snow deposition. Thus, the micro penetrometer hardness varies by a factor 4 within this stratigraphic layer. Layer B shows a consistent increase in hardness from 0.11 N to 0.8 N, an increase by a factor 8. Layer C entails two hardness increases and two decreases between 1 N and 8 N. Layer D shows a harder layer with a difference of 0.9 N to the layer below that does not appear in the hand hardness. This layer is captured in the surface image in Fig. 4. Layer E is similar to layer A, but at a higher hardness level. Layer F and layer G show the positional lag in the ram profile. Layer F has a gradual force decrease in the ram hardness and a sudden force decrease followed by a gradual force increase in the micro penetrometer hardness. Layer H is contradictory to the ram hardness. The homogeneous stratigraphic layer contains a gradual hardness increase by a factor 10, from 0.06 to 0.6 N. Layer I varies between 0.06 N and 0.6 N in the micro penetrometer where micro layers become apparent.

The following layers are from the spring profiles that contain wet layers that are very soft and refrozen layers that are on average much harder than the layers of the mid winter profiles. In layer J, a thin layer, which is softer by a factor 4, is missed in the hand and ram hardness. The ram hardness resolution is better in hard profiles than in soft profiles. Layer K contains two very soft layers, which are slush horizons. The SnowMicroPen measured faster than the drainage from the layer could occur to lower parts of the profile. After opening a wet snow pit the water can drain downward from saturated layers and re-strengthening can occur in these layers. This is shown in the systematically greater hand hardness than ram hardness in the spring profiles. Layer L is a rather homogeneous layer with small hardness variations.

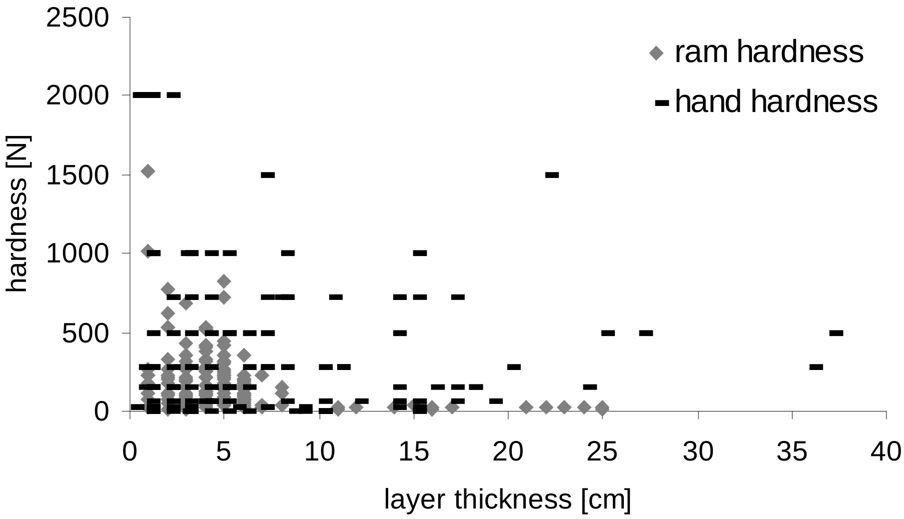

Fig. 6: The hand and ram hardness levels according to layer thickness. Ram hardness has a trend to more thin hard and thick soft layers, whereas hand hardness and thickness are more evenly spread.

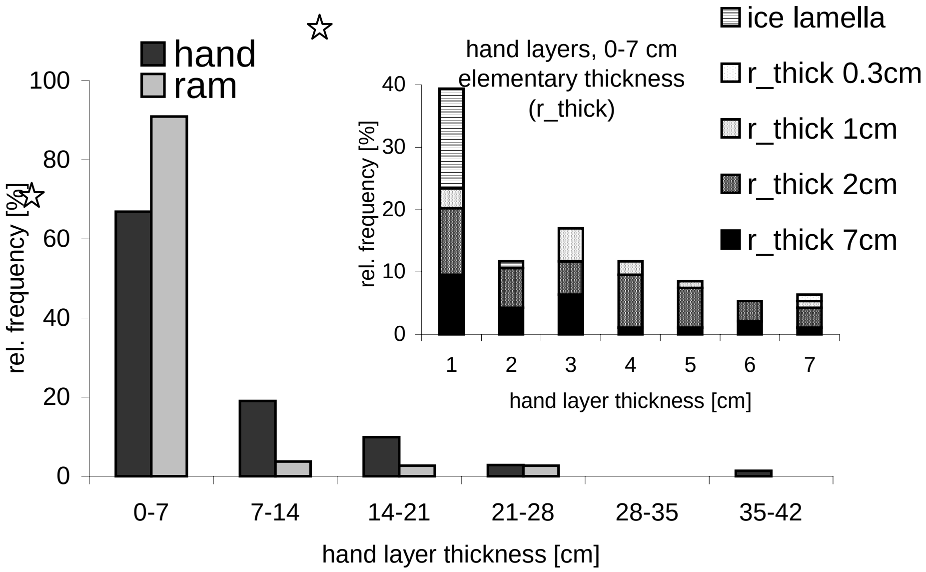

Fig. 7: Occurrences of layer thickness in ram and hand profiles. The inset shows the occurrences of hand layer thickness from 0-7 cm. The elementary thickness is the minimal layer thickness needed to perform the according hand hardness test (fist is 7 cm, 4 or 1 fingers is 2 cm, pencil is 1 cm, knife is 0.3 cm).

Classical stratigraphical layers in the investigated profiles show often an increase or decrease in micro penetrometer hardness, and are not uniform. Most layers, except a few layers in wet snow, show a strong fluctuation between 20 to 50 %. In wet snow the SnowMicroPen measures the profile fast enough to capture the properties before the water can drain from the wet layers.

6.4.4 Statistical comparison of ram, hand and micro penetrometer hardness

Systematic differences of ram and hand hardness from micro penetrometer hardness appear in shape of the cumulative probability curves, shown in Fig. 8. In the micro penetrometer profiles, about 50 % of all values are below 1 N. There is a higher probability to get high snow hardness values in a snow profile when measured by ram and hand hardness. The shape of the probability curves coincides only at layers above 500 N ram hardness. Ram and hand hardness have low resolution especially in soft snow. The lowest resolution of ram hardness is 10 N, which is given by the weight of the probe. Hand hardness resolves the soft snow only in two classes. The SnowMicroPen has equally high resolution in soft and hard snow.
The comparison of SnowMicroPen hardness and ram hardness is shown in Fig. 9. The SnowMicroPen hardness values in the recorded ram hardness intervals are illustrated as box chart where the 25th and 75th percentiles determine the box and the 5th and 95th percentiles determine the whiskers and the mean is indicated by a small square. The relationship is not linear, but exponential. Up to a ram hardness of 280 N the median micro penetrometer hardness is below 1 N. The hardness increase in the softer snow in the profiles is steeper in the measured ram hardness than in the micro penetrometer hardness. Due to the high spatial resolution of the SnowMicroPen measurement, which is at the grain and pore level, many values close to zero force are recorded. This also explains the positively skewed distribution of the micro penetrometer hardness values in all ram intervals up to 500 N ram hardness. In the extremely hard melt-freeze layers with ram hardness values above 500 N the distribution of micro penetrometer hardness becomes negatively skewed because the pores are greatly reduced in this snow. The micro penetrometer hardness signal in each ram hardness class shows greater variability because thin hard layers and soft layers are often missed by the ram test. The hardness variations in the micro penetrometer signal represent also the heterogeneity of the snow properties within the layers, such as the micro layering produced by wind influence during snow deposition.

6.5 Conclusions

Hand hardness, ram hardness and micro penetrometer hardness profiles are different records of the same snowpack. Each method is based on a different measuring process and has a different resolution in terms of space and hardness. The reproducibility of the SnowMicroPen field measurements is shown. From the comparison of the hardness profiles to highly resolved surface images of snow samples it results that the micro penetrometer profile captures the stratigraphic features more completely. Hand hardness profiles capture 80 % of the stratigraphic features and ram hardness profiles only 60 %. Besides the mean hardness, the hardness variation and micro textural information can be extracted from the SnowMicroPen force signal. No stratigraphic layer thinner than 1 mm could be found in the

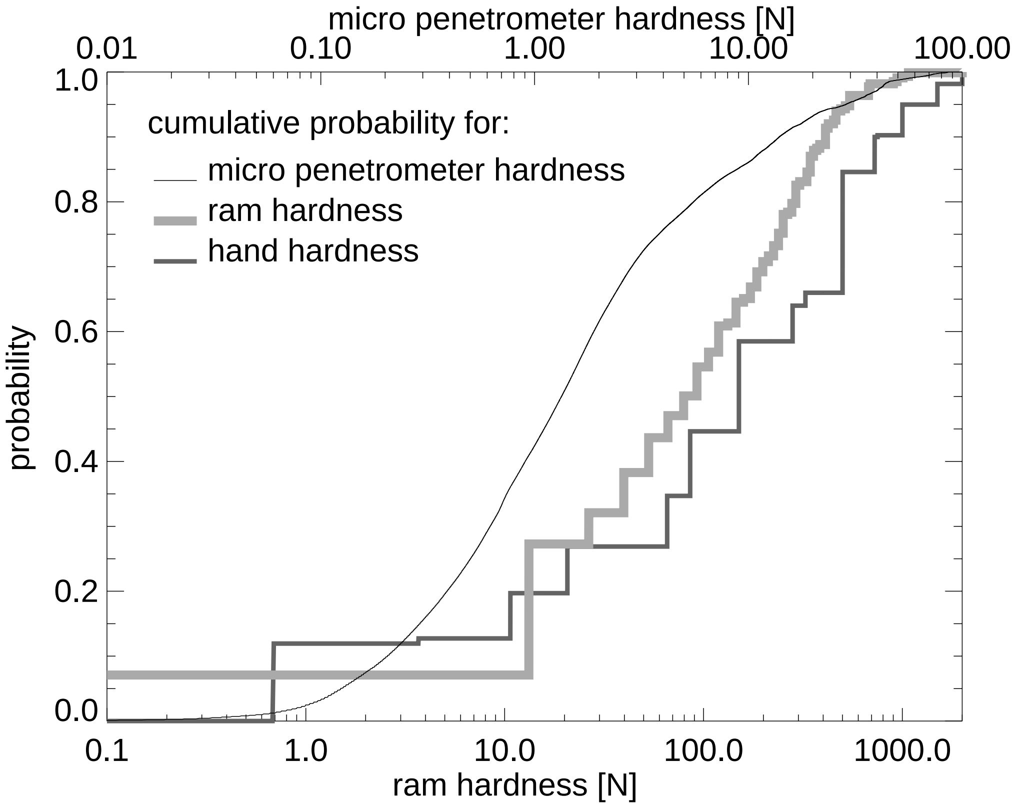

Fig. 8: Cumulative probability for ram hardness, hand hardness and micro penetrometer hardness on logarithmic x-axes. High hardness values occur with a greater probability in ram and hand hardness profiles than in micro penetrometer profiles.

surface sections. Thin layers occur frequently in snowpacks. It is not possible to measure the hand hardness of about 25 % of all layers with the hand hardness method, which is the basis for the snowpack stratigraphy. Classical stratigraphic layers show a strong fluctuation in all the profiles measured. In many cases a consistent trend over a large (factor 2-10) range in hardness was detected. This feature has not been measured, but was known to exist from translucent profiles (Good and Krüsi, 1993). The comparison of micro penetrometer hardness and hand hardness is shown in Fig. 10. The relationship is similar to the ram hardness comparison but the hand hardness increase in soft snow is even steeper. In soft snow the hand hardness measurements are even higher than the ram measurements. Hand and ram hardness tests measure snow, and especially soft snow, systematically harder than the micro penetrometer.

In the ram and hand profiles a considerable number of stratigraphic elements were missing, such as thin hard layers and soft layers. The hand hardness has better vertical spatial resolution than the ram hardness. The hardness differences at layer boundaries are correlated between hand and micro penetrometer hardness. However, the hardness resolution of the hand test is limited by the elementary area of the measuring device. Customary solutions are known but not always applicable and not documented. Hand and ram hardness tests measure snow, and especially soft snow, systematically harder than the micro penetrometer.

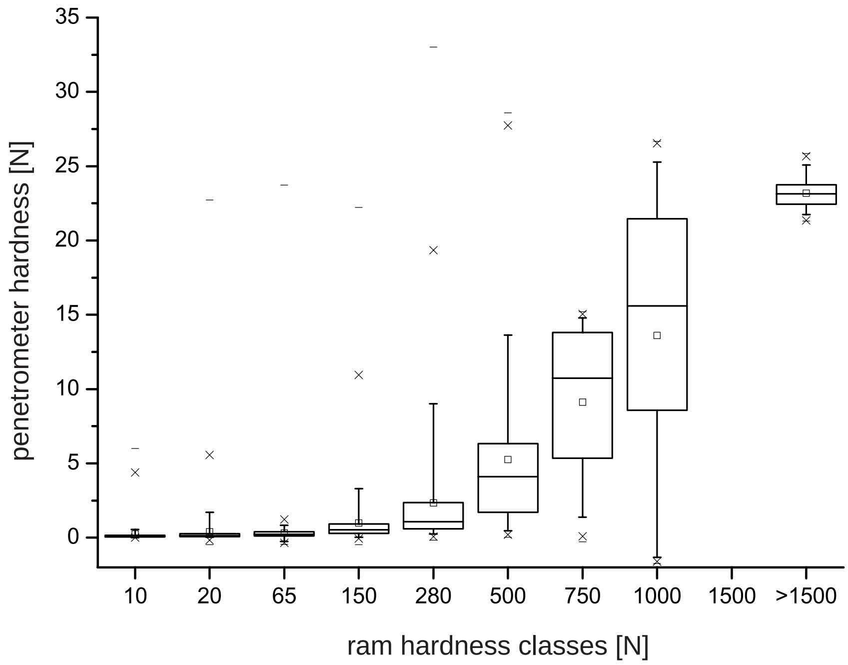

Fig. 9: Comparison of micro penetrometer hardness and ram hardness. Micro penetrometer hardness values in the respective ram hardness intervals are summarised in box charts. The box is determined by the 25th and 75th percentiles and the whiskers are determined by the 5th and 95th percentiles. The mean is indicated by the small square. First and 99th percentile are drawn as x and minimum and maximum as dash.

Classical stratigraphic methods should be applied with great care to quantitative comparisons. This study shows that methods with high spatial and hardness resolution are necessary to detect the subtleties of the stratigraphy in the natural snow cover. Other high-resolution methods (surface sections, 3-D reconstruction, translucent profiles) are more time consuming than the micro penetrometer. The complexity of a snowpack measured with the SnowMicroPen is surprising. It will be necessary to develop improved algorithms to extract all the information and to quantitatively classify layers and layer boundaries in a micro penetrometer hardness profile. Process studies will get a much more detailed input from SnowMicroPen measurements and model runs can be compared with more complete stratigraphic data. The classical concept of layers, which is valuable for operational avalanche warning purposes must be used with caution when applied to the simulation of snowpacks (Lehning et al., 2001) and the simulation of snowpack stability. Further physical interpretation of the micro penetrometer hardness will be a next step. The correlation to bond properties from stereological analysis will enhance the signal interpretation. This could widen the use of micro penetrometer hardness measurements. A further step is to correlate the micro penetrometer hardness and texture of failure layers to the layer stability.

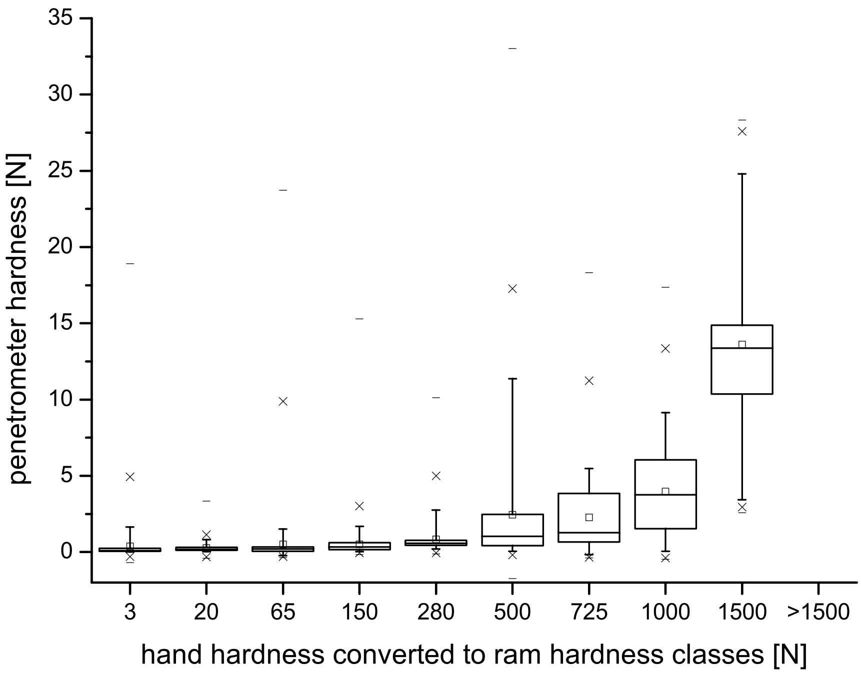

Fig. 10: Comparison of micro penetrometer hardness and hand hardness. Micro penetrometer hardness values in the respective ram hardness intervals are summarised in box charts. The box is determined by the 25th and 75th percentiles and the whiskers are determined by the 5th and 95th percentiles. The mean is indicated by the small square. First and 99th percentile are drawn as x and minimum and maximum as dash.

6.6 Acknowledgement

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Chapter 7

Synopsis

The textural and mechanical properties of mountain snowpacks can be objectively captured with the SnowMicroPen (SMP). The probe delivers physically relevant parameters and captures the variations inherent in natural snowpacks. A method for the investigation of the small scale spatial variability is developed. The signal interpretation is developed for the vertical variability of mountain snowpacks. The classical perception of a layered snowpack is called into question by the results of this investigation. The research results, the incurred problems and the relevance of the findings are summarised. Finally, an outlook on the next research steps and on potential practical applications is given.

- A comprehensive study of the developments in snow stratigraphy is carried out to follow the different paradigms that have existed. Apart from a unique historical review, this study is also an evaluation of the directions that have been taken and are taken today in snow stratigraphy. It points to directions that yield the physically most relevant representation of the complete snowpack. It also shows valuable methods developed in the past that were not followed. The most promising approaches are:

  - The combination snow textural and snow strength analyses,
  - understanding and quantifying surface properties,
  - methods that capture the spatial snowpack variability with its relation to processes such as water flow or avalanche formation, and
  - methods that help establish scaling laws in the spatial and temporal variability of mountain snowpacks.

- Different signal analysis algorithms for the interpretation the SMP force-distance signal are presented. Hardness and texture can be resolved from the signal with high resolution. The comparison of classical snow profiles, translucent profiles and micro penetrometer profiles shows that micro penetrometer profiles and classical profiles are not directly comparable. Snow layers and layer boundaries do not appear with equal clarity in the SMP profile. The variations in the snow stratigraphy retrieved qualitatively from the translucent profile were reproduced in the SMP profile.

- The texture index, a classification algorithm calculated from the SMP force signal, is successfully applied to verify numerical snowpack simulations of laboratory experiments on temperature gradient snow metamorphism. However, the comparison of measured and modelled snow properties of field snow profiles reveals systematic deviations. Model formulations and field methods for measuring snow properties need to be improved. The texture index calculated from SMP measurements captures detailed stratigraphic elements, such as thin crusts. It is suggested to complement the classic stratigraphy with SMP profiles in order to improve the verification of snowpack simulations.

- Snow stratigraphy and physical snow properties of natural and artificial ski slopes can be characterised from high resolution micro penetrometer measurements. SMP hardness is correlated to density and thermal conductivity. It is assumed that the increased density of the processed snow on ski slopes leads to soil cooling or even to soil frost. However, for the investigated study site, mechanical snowpack processing had no influence on the soil thermal conditions. A differentiated analysis showed that soil properties, snow height and average air temperature compensate the effect of the increased thermal conductivity of snowpacks on ski slopes. A tool to model the properties of the processed snowpack is successfully implemented in a soil model of energy and heat fluxes.

- Systematic measurements of the vertical variations of the textural and mechanical snowpack properties were carried out in natural snowpacks. The analysis and the comparison to highly resolved images of the snow profile shows that the quantification of the textural and mechanical properties of mountain snowpacks is now possible. Layers as thin as 1 mm are found in the images. The comparison of classical and micro penetrometer hardness profiles shows that only the highly resolved SMP profile delivers a complete snowpack stratigraphy and resolves the snow hardness and texture at the micro scale. It is also shown that classical hardness tests systematically overestimate the snow hardness.

- During the systematic spatial variability (cm to m) field measurements problems arose with the SMP. The signal analysis showed a non-linear distortion of the results. The usually strong vertical temperature gradient in natural snowpacks caused a strong drift in the force sensor, which required additional calibration. This effect was especially severe in soft snow, which is predominant in natural snowpacks. Laboratory tests were carried out and a rectification algorithm was developed. The calibration of the SMP before and after the field season 2000/01 deviated significantly from the laboratory constants. Further laboratory tests revealed that excessively tight fitting and small lesions on the tip-shaft of the SMP caused this. The latter effect is very difficult to rectify numerically because of its highly non-linear nature. It is impossible to retrieve the micro structural properties from the field dataset in an unambiguous manner. However, the situation was improved during the summer 2001 by replacing the sensor. The drift of the new sensor is 20-times smaller than that of the old one. Tests in the field were successful and the alterations allowed continuing the investigation of the micro structural properties and their spatial variability. In equi-temperature snow conditions, which were present in laboratory experiments and in field measurements of spring snowpacks (spring 2000), both having had equi-temperature conditions, the original force sensor did not have any drift and the interpretation of these measurements is valid. Systematic measurements of the lateral snowpack variability were carried out during the winter 2001/02. The analysis and the development of classifications algorithms are not finalised and therefore not documented in the thesis.

This work shows that measurement methods with high spatial resolution are necessary to detect the subtleties of the natural snowpack variations. Other high resolution methods (surface sections, three-dimensional reconstruction, translucent profiles) are more time consuming than measurements with the SMP. The snow stratigraphy and snow properties gained from SMP measurements are very complex. It will be necessary to develop improved algorithms to extract all the information present in the SMP signal and to further develop hierarchical classification algorithms for snow micro properties, the properties of layers, layer boundaries, horizons and surfaces from the SMP profile. Process studies will gain from a much more detailed input from highly resolved measurements and snow cover models can be compared with more complete stratigraphic data. It is suggested to implement the profile interpretation of SMP measurements in operational avalanche warning services. A test in this practical application will show in how much the new method can directly improve avalanche prognosis. The classical concept of layers, which is valuable for the bulk characterisation of the snowpack properties, must be used with caution when applied to the simulation of seasonal snowpacks and snowpack stability. The recent developments in snow stratigraphy challenge the general assumption of a snowpack consisting of discrete layers with homogeneous properties. The classical qualitative and gross sampling could be replaced by a quantitative and detailed sampling, giving more insight into the mechanical, thermal and hydrological behaviour of snow.

Further physical interpretation of the micro penetrometer signal is a next step. The correlation to bond properties from stereological micro tomography analysis will enhance the signal interpretation. This could widen the use of micro penetrometer measurements. A further step is to correlate the micro penetrometer hardness and texture of failure layers to the layer stability. Besides the physical interpretation, the analysis of the spatial variability at different scales will reveal the determinants of the spatial snowpack variability. It will also show whether scaling laws can be developed. A combined physical and spatial analysis of the mountain snowpack will potentially advance our knowledge about this highly complex spatio-temporal system.

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