Diplomarbeit

Titel der Diplomarbeit

Investigation of Water Storage and Flow Processes in Snow, Firn and Ice:

A glacial-hydrological Investigation on the Alpine Glacier Goldbergkees in the Sonnblick Region (Hohe Tauern, Salzburg, Austria)

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View from the *Hoher Sonnblick* (3106m), line of sight: South-East, *Source: own picture, April 2006*

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Preface and Acknowledgment

In the course of a field tutorial organized by the Department of Geography and Regional Research (Universtiy of Vienna) and the Central Institute for Meteorology and Geodynamics (Zamg) arose the idea to investigate special aspects of water storage and discharge behavior of the Golbbergkees glacier in the Austrian Alps. Mainly, Dr. W. Schöner (Zamg) and Dipl. Ing. G. Koboltschnig (University of Natural Resources and Applied Life Sciences; BOKU) adverted to the lack of information regarding storage and discharge behavior as well as en- and subglacial channel characteristics for the Goldbergkees.

The present study should therefore emphasize on the investigation with coloured and fluorescent dyes in order to investigate storage and flow characteristics in the summerly ablation period of 2006.

The gathered information should complement results for the project SNOWTRANS (Transformation of observed and computed ice and snowmelt data to ungauged basins. Even if the project was closed during the elaboration of the present thesis, results from this study contributed as vice versa.

Preparation in terms of theory, methods and equipment was done in spring 2006. Field tests were conducted in different field campaigns from June to October 2006.

The major part of the present study has been supervised by Dr. Schöner W. (Zamg), DI Koboltschnig G. (BOKU), Univ. Prof. Dr. Glade T. (Department of Geography and Regional Research; University of Vienna) and Univ. Prof. Dr. Hofmann T. (Department of Environmental Geosciences; University of Vienna).

The following departments were involved in the outcome of this thesis:

- Central Institute for Meteorology and Geodynamics (Zamg)
- Department of Geography and Regional Research (University of Vienna)
- Institute of Water Management, Hydrology and Hydraulic Engineering (IWHW, BOKU)
- Department of Environmental Geosciences (University of Vienna)

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Moreover, I would like to thank my study fellows Rebecca Mott (University of Vienna, now SLF-Davos), Gernot Michlmayer (BOKU, now ETH-Zürich), Daniel Binder (Vienna University of Technology) and Benhard Hynek (Zamg) for the mental and in field support, as well as my family and closest friends.

ABSTRACT

The diploma thesis „Water storage and flow processes in snow, firn and ice: A glacial-hydrological investigation on the alpine glacier *Goldbergkees* in the Sonnblick Region (Hohe Tauern, Salzburg, Austria) "analyses various processes of water transport in glaciated system. It measures in particular cycles and processes of water transport in different glacier sub-systems in the ablation season of 2006 and are placed it into a temporal context. The glacier itself is seen as the element of retardation in run-off as storage of water. The spotlight is on measuring retention times and quantities as well as interacting compounds.

In a first step investigations on terrain characteristics and the selection of representative experimental positions were done.

Possible test sites were identified by field inspections and careful (GIS).

The application of coloured dye on snow and firn uncovers near surface water transport processes. In the initial ablation season of 2006, infiltration processes and the progress of the melt front were investigated. Results showed a slow propagation of meltwater into the layered snow cover because of a preferentially slope-parallel movement. Water infiltrated faster through failures in the layered structure and vertical flow fingers in comparison to the surrounding areas. It resulted in a stepwise intrusion of meltwater.

The main part of the study emphasized on fluorescent dye tracer measurements in order to gather information about retention times of water in the different parts of glacier drainage system. By applying fluorescent dyes and their detection on selected points of interest, flow times were measured. Dye measurements under snow on firn showed significantly reduced flow velocities in comparision to later en- and subglacial flow measurements.

To measure the englacial flow of water, again fluorescent dyes were used. From measured data, flow rates under snow respectively on firn and ice, as well as flow velocities in the main englacial and subglacial channels were derived. Measured passage velocities were between 75 and 133 m h-1 for englacial flow. It could be compared to results from melt/rainfall-discharge analyses showing overal lag-times of 6-7 hours for July and 4 hours for September investigations.

Meteorologic and hydrographical data from the Sonnblick *Observatory* resp. the nearby gauge station were used to support analyses and results.

The comparison of outputs from different methods provides new information regarding flow processes and hydrological storage characteristics of a glaciated alpine drainage area.

KURZFASSUNG

Die Diplomarbeit “Water Storage and Flow Processes in Snow, Firn and Ice; A glacial hydrological Investigation on the Alpine Glacier *Goldbergkees* in the Sonnblick Region (Hohe Tauern, Salzburg, Austria)“ untersucht die unterschiedlichen Prozesse des Wassertransportes im Gletscher. Ziel ist es die Abläufe und Prozesse des Wassertransportes am Beispiel der Ablationsperiode 2006 in den unterschiedlichen Teilbereichen des Gletschers zeitlich zu messen. Dabei wird der Gletscher als Einflussgröße der zeitlichen Abflussverzögerung sowie der temporären Speicherung von Wasser untersucht. Die Verweildauer, die Menge des Wassers und deren beeinflussende Faktoren stehen im Mittelpunkt des Interesses.

In einem ersten Schritt wurden Untersuchungen zum Gelände, zur Wahl repräsentativer Teststandorte und zum idealen Untersuchungszeitraum durchgeführt. Mit Hilfe von Geländebegehungen und detaillierter GIS - Analyse wurden mögliche Mess-Standorte ermittelt.

Anhand von Markierungsstoffen wurden am Beginn der Schmelzperiode 2006 oberflächennahe Wasserbewegungen im Schnee sichtbar gemacht und ausgewertet. Die Markierungen zeichneten Prozesse und Wege des Schmelzwassers beim Einndringen in den Schneekörper auf. Durch die schichtartige Struktur des Schneekörpers kann das Wasser nur sehr langsam in die Tiefe einsickern und bewegt sich vorzugsweise hangparalell den Schichten folgend. Lokal begrenzte Brüche dieser Schichten und vertikale Fließkanäle förderten jedoch ein schnelleres Eindringen als im übrigen Schneekörper. Die Folge dieser abwechselnd schichtparalellen und vertikalen Wasserbewegung ist ein treppenartiges Vordringen des Schmelzwassers.

Der Fokus der Untersuchung lag allerdings auf der Messung von Fluoreszenztracerkonzentrationen, um Aufschluss über Verweilzeiten des Wassers in den verschiedenen Teilsystemen des Gletschers zu erlangen. Durch die Ausbringung von Fluoreszenztracern und deren Nachweis an ausgewählten Punkten konnten Fließzeiten bestimmt werden. Fließgeschwindigkeiten unter Schnee auf Firn waren erheblich geringer als in später bemessenen in- und subglazialen Kanälen.

Die Fließgeschwindigkeiten betrugen ca. 75-133 m/h und konnten mit Charakteristika von Abflusskurven bei Regen- und Schmelzereignissen verglichen werden. Für letztere wurden Verzögerungsintervalle von 6-7 Stunden im Juli bzw. 4 Stunden im September gemessen; ein Indiz für die Abnahme der Speicherfähigkeit des glazialen Systems.

Die Verknüpfung der Ergebnisse lieferte spezifische Aussagen über Fliessprozesse und hydrologische Speichereigenschaften eines alpinen vergletscherten Einzugsgebietes.

1 Introduction and Structure

1.1 History of Glaciologic Investigations on the Goldbergkees

Since 1886 the *Goldbergkees* takes a centre stage in the investigation of glaciers in Austria. With the erection of the Meteorologic *Observatory* on the *Hoher Sonnblick* peak (3105m) the basis for detailed research and observation was established. The *Observatory* provided the possibility to collect meteorologic data in situ and monitor the results by permanent observer. Therefore we have access to long and continuous series of climate temperature measurements. Not till then scientific work was possible and gave birth to valuable meteorologic, climatologic, hydrologic and glaciologic investigations.

In 1896, the geographer Albrecht Penck was the first to map the three main glaciers near the *Observatory (Penck, 1897, as cited in Koboltschnig, 2006, p.7).*
In 1909, a new method of mapping was applied to the glacier of *Goldbergkees*: the terrestrial photogrammetry, which mapped the glaciated site in a scale of 1:10,000 *(Hübl, 1912, as cited in Böhm, 1986).*
Until the late 1960 mostly investigations on radiation balance were made. In 1971/72 in the course of the hydrologic decade of the UNESCO Austria participated with a study on the thickness of the *Austrian* glaciers (including Goldbergkees, Würenkees, Kleinfleisskees). *(Brückl & Bittmann, 1977, as cited in Koboltschnig, 2006, p. 8).*
Since 1983 the most important studies carried out in the Goldberggruppe are direct mass balance measurements. Starting with the *Wurtenkees* in 1983, followed by the *Goldbergkees* (1987) and the *Kleinfleißkees* (1999) continuous direct measurements of accumulation and ablation are carried out. The continuous measurements of glacier balances allow us to better understand climate-glacier relationship.
One important component of the climatic and hydrologic measurement instrument network in the Goldberggruppe is the gauge station at the glacier tongue. In the first gauge station was installed at the outlet of the catchment area at the tongue of the *Goldbergkees*, followed by others at the Kleinfleißkees and the Wurtenkees. They provided exact information about the amount of discharged water emerging mostly of the glacierized areas above the gauge station.

Moreover they provide proper calibration data for hydrologic modelling of the discharge at the *Goldbergkees* by *Koboltschnig* (2007) and the project “Snowtrans” (Transformation of observed and computed ice- and snowmelt data to ungauged basins).
Due to the vicinity to the meteorologic *Observatory* at the *Hoher Sonnblick*, the measurements should be easily associated to the meteorologic conditions at the time of the

investigation. There are some other Austrian glaciers often being investigated. However a better meteorologic database is hardly found elsewhere.

On the basis of this premise originated the following work.

1.2 Structure of the Thesis

Chapter *"Review"* is supposed to outline the actual state-of-the-art in glacial-hydrological studies treating as well the *"Basics of Hydrology"* and the *"Basics of Glaciology".* The chapters emphasize the principles and basics relevant for this thesis study in order to not treat the complete glaciologial basics. This would go beyond the scope of this thesis.

The chapter *"Study region and Investigation Area"* covers the geographical frame in which the field investigations and the thesis are situated. The results extracted from this paper must be regarded within this spatial frame.

Chapter *"Methodology"* gives detailed information about the used methods for field investigation as well as pre- or postprocessing work. It discusses basic theories, formulae and principles of the applied methods. Why did I choose these methods? The chapter should design the consistent workflow for reaching the initially delineated aims.

Personally collected and received data (other party) was covered in an own chapter *"Data Base".*
For this thesis different data gained from tracer tests, photographic documentation, hydrometeorologic gauging stations and geographical data was used.

The chapter *"Results and Discussion*" is subdivided in the outputs from the different methodic approaches. The aim is to connect gained information, scientifical and personal knowledge to delineate the maximum of extractable information and to interpret it. Moreover, combining the various approaches should lead to additional information output.

"Summary and Conclusion" unifies all main statements to a single overall chapter in which main results get connected. As long as a glacial drainage system is very complex this chapter tries to outline systematic and consequential outputs.

"Perspectives" for further investigation, leakage of theories or methodology, constraints and autoreview or -criticism is done in the final chapter.

2 Review

The purpose of this chapter is to give a general overview of the actual scientifical state inside the frame of glaciological studies. What is the state-of-the-art in glacial –hydrological investigations?

Fundamental theories and knowledge of the principles of glacial hydrology are covered. Hydrological processes on glaciers are rather complex. Pre-treatment of the basics facilitates access to the theme in such a way to enable better understanding.

2.1 Basics of Glaciology

"Glaciers may be defined as an accumulation of ice and snow that moves under its own weight in response to gravitational force" *(as cited in Singh, 2001, p. 448).*

The formation of glaciers at any location is related to the existence of wintery or summery solid precipitation, its accumulation and methamorphism to glacier ice. Glaciers can only exist if accumulation rates is greater than ablation. This means that the amount of fallen snow is relatively greater to the amount of water being lost due to melting and evaporation. It also means that at least on some parts of the area snow is surviving the annual cycle and then transformed into firn.

It assumes that temperatures are relatively low and favour the occurrence of solid precipitation. Low temperatures are frequently in high latitudes eminently in the so called Polar Zone above the 60° mark. But they can generally exist in any latitude presumed that climatic conditions are adequate. In lower latitudes such conditions can be found at high altitudes. In fact the existence of glaciers is significantly related to the altitude depending on the latitude. Glaciers in higher latitudes show a lower limit of existence. Though there are still differences mostly depending on the dryness of the climate and the annual distribution of precipitation. Wet and cold climatic conditions in winter favour the existence of glaciers, whereas under dry climatic conditions glaciers can only be found at high altitudes. Globally, a rising of the so called snowline from the poles to the equator is observable. The snowline for e.g. on the poles is at the sea level, in the Alps it is between 2500-3000 m and in between 5000-6000m in tropical regions *(Singh, 2001, p. 38).*

Glacier ice is formed by methamorphism of snow and firn. The compression of snow originates from its own weight as well as from superimposed snow masses when covered by new snow. The metamorphism of snow to ice runs significantly faster in temperate climates. Warmer temperatures in the ablation period favour faster transformation processes. At the end of the process glacier ice reaches a density of typically 820-900 g/cm³.

The precipitation is the intrinsic source of accumulation. But there are secondary wells for the accumulation. Avalanches can play an important role in the formation of glaciers. Moreover the wind drift of snow erodes snow on ridges and exposed areas and deposes it on areas of lower wind velocities. Those processes are not to be neglected. It is generally accepted that some glaciers are strongly dependent on these processes.

The most of the glaciers in mountainous regions, about 50% are situated in the Himalayan region *(Bahadur 1992, as cited in Singh, 2001).* With exception of the Antarctica the greater number of glaciers is found on the northern hemisphere.

Glaciers can vary greatly in size and area. There is no lower limit which defines the minimum size for a glacier classification, even if glaciers under 1 km² hardly attract attention. They vary in thickness to. Glaciers smaller than 1 km² often have only thicknesses of a few tens of meters, whereas huge valley glaciers and ice sheets can have thicknesses of more than 100m and even more than 1000m. The Antarctic ice sheet has thicknesses of more than 4000m *(Singh, 1997).*

2.1.1 Properties of Snow and Ice

2.1.1.1 Transformation of Snow to Ice

Solid precipitation in form of snow is the basic requirement for the formation of ice. Snow crystals are deposited on the ground. Independent of their varying forms they immediately get transformed. Due to free energy differences on the snow crystal forms, water molecules begin to move from the tips of the crystals to the gorges or mostly the center of the structure. The laws underlying this process are thermodynamical and consist in the fact that the free energy of molecules in a system (represented in this case by the single snow crystal) tends to a minimum. This process results in a general rounding of the crystal structure and a transformation towards more spherical forms of crystals. The snowflakes with high surface energy become rounded and are developed into rounded snow with lesser diameter and surface energy. The warmer the temperatures the faster this process is taking place *(Paterson, 1994, p.20)*. Temperatures near the freezing point favour the migration of molecules on the crystal surface, as well as to the air and back in form of refreezing. Therefore transformation is decelerated in Polar Regions. An important side-effect of this process is a decrease in air volume of the snowpack leading to an increase in density. Moreover the snowpack is affected by a compaction. In general we identify this change in

morphology as the snow metamorphism and more specifically as equitemperate or destructive metamorphism *(Singh, 2001).* Bigger grains grow in expense of the smaller.
Due to surface melting water is percolating into the snowpack. If the snowpack is equitemperate it will get to the water table. Introduction of water into the snowpack leads to abased capillary water pressure which accelerates the crystal decomposing process. Water content resulting in water regimes is subdivided as follows: the capillary regime (water content under 10% by weight), the pendicular regime (water content less than 14% by weight), and the funicular regime (water content over 14% by weight); (Fig. 1)

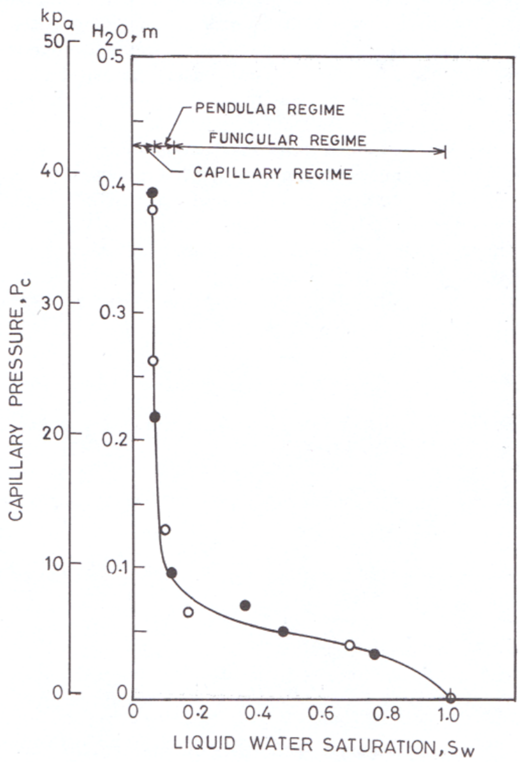


Fig. 1: Capillary pressure in wet snow as a function of water pressure, hollow points snow with mean density of 590 kg/m³, black filled points: snow with a density of 550 kg/m³; *(after Colbeck, 1973, as cited in Singh, 2001, p.211).*

Especially in seasonal transition periods as spring and autumn subzero temperatures by night lead to refreezing in the snowpack or on the snow surface as well. The refreezing can entrap water and fill the pore spaces between the grains. The consequence is the creation of bonds in other words a melt together of the grains and a further increase in relative density.

Snow lasting over the summerly melt season exposed to the processes described above is generally identified as “firn”, even when there is no clear division from snow. The identification comes mostly of the relative densities as listed in **Tab. 1**.

Tab. 1: Typical densities of different natural phases of water in g/cm³

New snow (shortly after deposition)	50-70
Damp new snow	100-200
Settled snow	200-300
Wind packed snow	350-400
Firn	400-830
Very wet snow and firn	700-800
Glacier Ice	830-917

The distinction of firn and ice in contrast is relatively clear. It is distinguished by the fact that ice does practically not possess air passages. In the course of the transformation the air filled capillaries and pathways get sealed off. Though there can be still air because natural ice has a small amount of enclosed air bubbles.

2.1.2 Spatial Properties of Glaciers

A glacier can be subdivided: Accumulation zone and ablation zone with the glacier tongue.

2.1.2.1 Accumulation Zone

The accumulation zone is in general the upper part of a glacier. Snow falling in winter does not melt completely in summer. In consequence total mass balance in this zone tends to be positive, which means accumulation exceeds ablation.
Accumulation of snow can be ascribed not only to solid precipitation but also to wind drift and to deposition due to avalanches.
A cross section reveals a layered composition. On top lies the seasonal snow cover. During the ablation season this snow cover melts partly. Warmer summers result in greater melt

quantities. Underneath, there can be found a firn layer, composed by “old” snow from the preceding meteorologic years. The old snow layer is lying on previous firn layers or bare ice. The accumulation zone is characterized also by the fact that there are typically no surface melt streams. Water coming from precipitation or melting is percolating into the snow- firn aquifer until it reaches the saturated snow layer. In most of the cases we can assume that there are three further subdivisions of the accumulation zone: a dry snow area, a percolation or transit area and a soaked (saturated) area. One could assign these areas horizontally or even vertically (*Müller, 1962 as cited in Paterson, 1994, p. 10)*.

The dry snow is the topmost part of the accumulation zone. For polar glaciers or glaciers starting above a potential melting line, melting is hardly observable. The mean temperature is under 0° Celsius.

Moving further downward and reducing absolute height there is the percolation area of the accumulation zone. With rising increasing irradiance temperatures are raised. Higher energy inputs into the snowcover evoke melt and cause water to infiltrate into the snow body. The mean particle snow size is greater because of the processes described in chapter 2.1.1. Due to refreezing of the passing water it is typical to find ice lenses, ice layers and snow clusters. Even though the boundaries to the soaked – water saturated zones are not clear. The water table in the snowpack can fluctuate highly during the year or even the day. Better transport capacities in the late ablation season turn the drainage to be faster than normal. Moreover increased solar radiation or warmer temperatures cause higher melt rates which can exceed the transport capacities, causing the water-table to rise.

The soaked zone is typically water saturated lies on bare ice, firn or super imposed. These soaked areas are increasing during the ablation season. The highest existing line is ascending with rising temperatures. The temperature inside the soaked zone is 0°C *(Paterson, 1994, p.11)*

The soaked zone is only water saturated if there are no easy drains into the ice body. Therefore we can assume that the soaked area is a discontinuous layer which fluctuates in time and thickness.

The accumulation zone is bounded by the firn line where a glacier is quantitavely stable. In glaciology it is defined also as the equilibrium line.

The firn line a glacier has a balanced mass budget with an equal accumulation and melt rate. The equilibrium line a reasonable indicator for the mass balance of a glacier. The equilibrium line tends to react sensitive to meteorologic or climatic changes. It tends to a stable position under stable climatic conditions. The firn line can be made out on the climax of the ablation season by the distinction of old snow and ice *(Singh, 2001, p.455-484).*

2.1.2.2 Ablation Zone

The areas were ablation exceeds accumulation is called ablation zone. The zonal net mass budget is negative. Due to gravitational forces the ice body is constrained to move downward. This movement brings parts of the glacier into lower elevations where melt exceeds snowfall.

The equilibrium line divides the accumulation zone from the ablation zone. Over a medium time scale the equilibrium line stays in the same absolute position even if the glacier is moving downward underneath.

2.1.2.3 Glacier Tongue

For temperate glaciers, the glacier tongue lies totally inside the ablation zone. The glacier tongue is the maximum extend of a glacier and gets its name because of the typical tongue form. As a result of the viscosity ice is normally a cohesive mass, typically with rounded edges and a rounded end. In the majority of cases we can make up a glacier snout at the longitudinal edge of the glacier tongue. It represents the exit of water out of the englacial hydrologic system. The glacier snout reacts to mass balance changes.

2.1.3 Glacier Movement

Mountain glaciers are “nourished” by accumulation of falling snow, wind drifted or avalanche deposited snow. The transformation of snow to ice replenishes the ice body of the glacier. Because of its own weight and the related effect of gravitational forces it drives the glacier body to move. Due to high stresses and the specific deformation properties of ice, describable as a visco-plastic body, the glacier is moving as a cohesive mass. To a certain threshold force ice can resist and behaves like a plastic manner. Time is playing an important role in deforming ice bodies. Ice crystals can rearrange in position following pressure forces (pressure deformation or sintering). If the acting shear stresses are not exceeding a threshold value, ruptures are not occurring. The ice body gets deformed but behaves as a unique cohesive body. Only by exceeding the shear stresses i.e. by a faster flow over obstacles, or by flowing over bigger obstacles, ruptures are observable. It can be easily made out on glacial ice falls, where greater deformation forces in shorter time periods lead to a cracking and to the formation of crevasses.

Though, the acting forces and stresses in a glacial body are rather complex and are not completely understood by now.

Basically there are two different areas in glacier flow: the accumulation area and the ablation area already treated of in chapter 2.1.2.1 and 2.1.2.2.

However, the glacier flow leaves the profile of a glacier remain relatively unaltered to the year before, despite the completely different mass budgets in the different zones. Because of gravitational forces acting on the enormous weight, the glacier is flowing from the accumulation to the ablation zone. Nevertheless the amount passing any cross-section of the glacier remains stable. The amount of ice flowing through the cross-section at the firn line must be the annual total amount of collected snow in the accumulation zone. The amount flowing through these cross sections is 0% of the annual budget at the longitudinal edge (glacier tongue, snout) and 100% on the firn line. Flow velocities would behave in the same manner if terrain properties were constant for the hole area. We could observe the greatest flow velocities at the equilibrium line *(Singh, 2001, p.460).*

Paterson (1994) reveals that velocities are increasing in fourth power with thickness and third power with slope angle. Paterson also calculated that deformation is running ten times faster at temperature around the freezing point than at -22°C *(Paterson, 1994, p. 235)*.

2.1.4 Classification and Inventory

The sense in setting up a classification of worlds glaciers is to consolidate their similarities. It should get the understanding of the associated characteristics and processes easier.

For this reason there are different distinguishing criteria.

One criteria is the thermal characteristic of the ice. Polar glaciers for e.g. are so called “cold” glaciers. The temperature of the glacier ice in average is considerably lower than for temperate glaciers. In fact the word “temperate glacier” anticipates the temperature of the glacier ice around the melting point or slightly lower. Only the near surface ice layers can be very cold due to cold surface temperatures. But the thickness of these layers is very mean, because of the low temperature conductivity of ice. Especially at the glacier base, where geothermal heat is giving its energy to the ice, polar glaciers can be temperate in some zones as well.

The characteristic of temperate and polar glaciers vary significantly. The rheological properties of glacier ice change with different temperature. Higher temperature favour deformation rather than breaking of the ice body when moving.

The most apparent criteria is topography, which is related to the ground conditions but above all to different meteorologic and climatic conditions heading to different accumulation and ablation.

Singh (2001) gives three different types of glaciers:

Ice Sheets, and ice shelves:	
	Ice sheets are continuous areas of great extent. Small ice sheets are classified as ice caps but do merely differ in the processes underlying its formation. They form when precipitation, and so accumulation exceeds denotatively mass losses due to melting. Moreover terrain characteristics must not excess a certain steepness not to build a valley glacier. The form is circular or plane more than stretched. If they cover a whole continent as in the case of the Antarctica they can be described as Continental glaciers. The movement does not particularly show a certain direction, rather than propagation in all directions starting from the highest point or the point with the most mass accumulation. Examples for the existence of Ice Sheets and ice shelves are the Antarctica, Greenland, or Iceland.
Valley glaciers:	
	Valley glaciers are the most common form of glaciers worldwide. This type of glacier is characterized by its longitudinal shape and its movement, often following the valley or even eroding them by themselves. The glacier can consist of accumulation zones, ablation zones, moraines (middle, side, end), tongue, crevasses zone and side glaciers. Typically they are the continental or alpine form of glaciers and are very common in the mountainous regions of the temperate northern hemisphere.
Piedmont glaciers:	
	A piedmont glacier is a hybrid between an ice sheet and a valley glacier. Emerging from a steep mountainous area as a valley glacier it is spreading on a fore lying plane and forming an ice sheet. E.g. In the mountainous regions of the Antarctica they are a very common phenomena.

Certainly there are some intermediate forms but not to be mentioned extensively in this work.

To have a better understanding and to get accurate information of the ice covered areas it is of great importance to observe extensively. It is important to document changes of glacierized regions. Especially changing climatic conditions have effects on the mass balance, on area, surging or retreating effects.

There are mainly tree different methods to monitor glaciers.

- *Ground survey:*

The method consists of in situ observations on the basis of topographic maps and cartographic means. The method implies measurements and observations of height, distance and area, firn lines, ablation areas, tongue areas, channels systems on the ice surface. Moreover, also geomorphologic processes are mapped. In addition snow, firn and ice sampling are common observation methods.

The precision though can vary depending on the data basis an on the observer itself. Furthermore the ground survey is rather time and workforce consuming.

- *Aerial photography*:

To improve precision and time in observation glaciers, aerial photography is carried out. The aerial photography does bring good results as long as done in the adequate moment.

- *Satellite observations:*

For regions with bad accessibility the only way to observe glaciers is the satellite imaging. Modern satellites can get resolutions up to 10-20 centimeters and are of the same quality as aerial photography. Moreover they often can document glaciers with different equipment as for e.g. IR-imaging. Problems can occur because of high reflectance of snow or ice.

All the gathered data is collected in an inventory which contains data on the location, elevation, area, width, thickness (if measured), withdrawal patterns, as well as other data.

2.1.5 Mass Balance Characteristics of Glaciers

Mass balance investigation on glaciers is an important issue monitoring the responses to trans-annual meteorologic or climatic variations. Mass balance is understood to be the difference between accumulation and ablation over a certain time period. In the Alps a glaciological or hydrological cycle starts with the beginning of accumulation in October and ends with the end of the ablation period in September. Consequently we refer to annual mass balance, net balance or annual budget. The mass balance only refers to mass changes and does not mean advancing or retreating of glaciers. Net balance changes are expressed in water-equivalent of ice units.

A very simple formulae is the one given below, where b_n is the total annual net balance, composed by the addition of b_w, the wintery balance and b_s the summery balance (negative).

$$b_n = b_w + b_s$$

For further understanding the formulae needs to be fragmented to its processes contributing to the total net balance. The basic process contributing to the wintery budget is accumulation(c). It is composed of solid precipitation *(P_{solid})*, refreezing free water *(P_{stored})*, condensation, wind deposited snow and avalanches.

$$Accumulation(c) = P_{solid} + P_{stored} + condensation + drift(positive) + avalanches(positive)$$

Summery balance is a product of ablation combining snow and glacial melt, evaporation, wind erosion of snow, avalanches triggered on the glacier and ending outside the glaciated area as well as calving.

$$Ablation(a) = melting + evaporation + drift(negative) + avalanches(negative) + calving$$

Paterson (1969), gives a combined formulae for b_n (net mass balance):

$$\boldsymbol{b_n = c_w + c_s + a_s = \int_{t1}^{tm} (\overset{g}{c} + \overset{g}{a})\,dt + \int_{tm}^{t2} (\overset{g}{c} + \overset{g}{a})\,dt}$$

,where cw is the accumulation in winter and c_s the accumulation in summer. Whereas as is the summerly ablation. Thus bn is the sum of the integral between the minimum of glacial mass at the end of the ablation season (t_1) and the maximum mass at the end of accumulation season (t_m) plus the integral of the maximum mass at tm and the minimal mass at the following year minima of glacial mass (t_2).

Singh (2001) identifies in this formula the different factors influencing the net budget:

$$Mass\ balance\ (b) = c - a$$

or

$$Mass\ balance = P_{solid} + P_{stored} - melting + condensation - evaporation + drift + avalanches - calving$$

Mass balance variations are mainly related to temperature and precipitation. Notwithstanding, small scaled meteorologic variations and differences can play an important role. The characteristics and the shape of the accumulation or ablations zone, as also its orientation can amplify special influences on the glacier balance.

Mass balance can be positive, negative or balanced. Small variations in mass balance can be within the normal range of glacier mass fluctuations, whereas long term trends suggest a climatic change or massive changes in budget affecting influences.

A series of positive balance years leads to an increase in thickness in the accumulation zone. With a certain time lag the glacier is responding to an increased supply of mass with the increase in flow rates. The glacier is automatically compensating the changed condition. The time in which glacial flow is responding depends on several parameters: amount of solid precipitation, average temperature of the ice (i.e. polar glacier, temperate glacier), glacier size, steepness (flow angle) and glacier shape. Increase flow leads to an altered glacier shape and size. Growth of ablation zones leads to increased ablation and does in turn balance the increase in accumulation.

A period of negative balance years results in a loss of ice, affecting both thickness and size. First of all thickness is decreasing because of a lack in solid precipitation. Flow rate is decreasing. Contemporaneously earlier snow-free areas on the glacier intensify ablation.

According to these principles it is evident, that positive balances result in glacier advance and negative balance years result in glacier retreat.

Budget quantities can be defined also for an area, which results in an average net balance expressed as a portion to the area of a glacier. For special purposes this can an interesting indicator allowing a comparison between glaciers of different size and location.

Thus

$$\bar{b}_n = B_n / A$$

where B_n is the net mass balance for a year and A the area of the glacier.

There are to be mentioned mainly three different methods: glaciological, hydrological or geodetic.

- Glaciological method: the method consists of snow sampling with pits and core drilling. The purpose is to define the water equivalent for the sampled snow through density determination. Even if it is a punctual sampling method it derives reliable data. Data can be easily enhanced by increasing the amount of sampling points. By inter- and extrapolating the derived water-equivalent it is possible to sum the total gained mass.

  Lost water-equivalent units are deduced of sampling stacks positioned in drilled hole in the ablation area. By surveying lost ice at the end of the ablation season water-equivalents can be calculated. Important for reliable results is the allocation of sampling stacks over the area, as well as the use material composition of these (stacks should be made of low conduction ratio materials). Results normally are plotted as net balance maps typically showing net balance isolines of water-equivalent units or meters of ice.

- Hydrological method: This method computes a total net loss or gain of water over a defined period. The method takes into consideration precipitation, run-off and evaporation leading to the output of net water storage or loss. Simple models provide only a total amount of water not considering aerial distribution. Modern hydrological modelling though can provide both two-dimensional distribution over the glaciated area and evolution in regards of the annual cycle. Kasser (1959), Tangborn (1966), and Stenborg (1970) propose hydrological methods as an adequate procedure to calculate the budget of glaciers. Generally it is expressed as:

$$b_n = P - R_t - E$$

or

$$net\ mass\ balance = precipitation\ \text{-}\ total\ runoff\ \text{-}\ evaporation$$

An important factor for precision is reliable data for precipitation in the watershed and good computation of actual and potential evaporation for the only partly glaciated watershed. In fact, the ratio between ice and ice free underground for the basin should be high. Moreover wind drift can distort results significantly.

- Geodetic (photogrammetry method): Extracting data of elevation changes out of accurate contour maps can be a reliable method for mass balance changes. Initially done with photographic means, nowadays satellite imagery, differential GPS or terrestrial geodesy provide even more accurate data. Though time expense can be relatively high comparing to lesser precise but faster hydrological methods. Imagery or photogrammetry survey has to be done at the end of the ablation season. The time of survey must be chosen carefully to avoid summing errors (summer balance, winter balance). The processes leading to the net budget are not quantified in this method. One can only generally quantify the amount of lost or gained ice mass over a year. Even if the method does not take into account the amount of gains or losses at a certain point we can assume that ice flow is balancing these local elevation changes (problematic only if flow velocities are changing abruptly).

Moreover depth measurements done with gravimetric methods, electrical resistivity method, seismic methods or radar applications can show long term changes in thickness of the ice masses and prevent data for budget measurements.

2.2 Glacial Hydrology

Nowadays glacial hydrology is often discussed in the context of global climate change and its consequences for certain aspects of environment. This could be the water balance of glacierized watersheds, changing geomorphologic attributes of watersheds or even the consequences linked to a sheer disappearance of glaciers.

Although often investigated glacial hydrology is still one main research field of hydrology and leaks of consistent theories and formulae. The highly changing circumstances and situations during glacial hydrologic investigations complicate comparability of the extracted information. Moreover they can only display a short time window in glacial behavior. Processes can change fast in diurnal, annual and interannual cycles and even though probable, similar conditions must not recur again. Regarding these circumstances there is still a scientific need for further investigations, as done in this paper. Only by collecting more data from different places under different conditions will brighten missing puzzles.

With a variety of methods research tries to calculate important glacial hydrologic parameters like discharge, total ice -water volume, peak discharge, reaction time of the hydrologic system *(e.g. Fountain and Tangborn, 1985; Hock, 1998).* They can provide important information not only for scientific purpose but also for civilian matters. Water supply in many regions is often dependent on the capacity of glaciers to store water and to release it when precipitation is missing *(Röthlisberger and Lang, 1987).* In addition glaciers can represent also a threat on the basis of its tendency to cause sudden outbursts of meltwater stored in the intra-channel system or in moraine-dammed lakes. Moreover, the quantification of glacial discharge is important for management purposes in hydroelectric facilities.

The changing climate represents a big challenge to deal with the rising problems of diminished total glacial volume and changed behavior of glaciers *(WGMS 1999 as cited in Schuler, 2002, p.2)*.

In the last decades mathematical modelling of glacial hydrology has become of rising importance. With changing climate conditions glacial hydrology will be affected as well. For this reason it is obvious that future dynamics of glaciers are linked to climate change. In order to deal with upcoming problems it is of great importance to develop calculable models. Whereas models for simulation of melt processes are quite accurate, water transfer models describing passage through the glacier ice deal with a black-box model. They base on linear reservoir models and that's why they exclude the dynamic nature and complex constitution of glaciers *(Baker, Escher-Vetter, Reinwarth, 1982; Jansson, Hock, Schneider, 2003; Mader and Kaser, 1994)*.

The investigation on glacial hydrology is not on a death end. Many aspects are neither quantified nor understood.

2.2.1 Basic Principles and Physics of Water Storage and Water Flow Processes

The watering through a glacierized system is a very complex interaction, induced by appropriate meteorological conditions and the physical components of the glacier itself. The three linked systems; snow, firn and ice are closely connected.

Water is stored in different matters within a hydrological system. There is to be mentioned the snowpack, firn, pools on the ice surface of different size as also crevasses, so called englacial pockets, the conduit network and even the basal sediment layer or moraine. The characteristics of the different elements alter markedly which results in different pathways and storage capacities leading to unequal transit velocities.

2.2.1.1 Storage and Movement in Snow and Firn

On and especially in snow, water percolates through the snowpack, due to gravitational forces. The contact between the loose or frozen snow grains causes friction to increase significantly. Therefore transitional dewatering is decelerated.

The processes passing in firn are characterized very similar. Contact zones of old snow grains cause additional granular surface as well as increased surface tension, which causes slowed passing velocities. Whereas ice layers often found in seasonal snowpacks are of great importance in snow they can not be made out in old firn layers. The compaction over a certain time (firn is glaciological defined as snow with an age of > 1 year) breaks up the ice layers. Certainly there are areas within a firn layer to have fewer or greater densities, but there can not be found well defined ice lamellae. The water in the firn layer percolates through widened veins.

The characteristics of the grained firn layer are often compared with the hydrological properties of a sandy soil layer. The grainy structure favors a slow and continuous depletion of water out of the aquifer if water inflow is stopped immediately. The properties in snow and firn layers do change over the year. Similarly to the moulin-conduit system in glacier ice, the continuous percolation of water leads to widened water channels. Pathways of relatively greater diameter grow in favour of the smaller capillary due to pressure gradients.

At the beginning of the melt season water pathways are hardly developed. Snow is covering the firn. Its water storage capacity is substantially higher than those of firn, due to lower average densities. Because of differentiated crystal structure and greater amounts of air inclusions internal friction is much higher in snow than in firn. Therefore it can hold greater quantities of water and delivering it much slower than firn (*Hooke, 2005, p.20).* With the progression of the melt season; the snow structure gets mostly cracked. The meteorologic conditions force snow crystals to simplify in regard of their form. Following the laws of snow metamorphosis the crystals are rounded and increase their diameter. The driving process is the so called melt metamorphosis also known as firnification. It is induced by above zero air temperatures, free water in the snowpack and refreezing by night. (*Gray and Male, 1981*). In the beginning of the melt season the snowpack is approximating the characteristics of firn. Distinguishing snowpacks from firn is considerably difficult with the progression of the season. Nevertheless there is still a difference in density and air volume. The melt water front is progressing through the snow/firn layer at the inception of the melt season. Investigations of melt front progress proofed that the outflow from not equi-temperated and therefore still partly frozen snow/firn packs is very poor and does vary diurnally and interdiurnally relative less than in the high melt season *(Fountain and Walder, 1998)*.

2.2.1.2 Storage and Movement in Ice (Englacial Water Transport)

Once reached the ice layer it can immediately enter the englacial transport network as also flow on the bare ice surface till it reaches an entrance to the englacial drainage system. The "entrance" can be a crevasse, a moulin or smaller hole in the glacier ice. It is accepted that water entered in the glacier ice itself, flows through a series of linked conduits and passages, as well as linked cavity systems or permeable sediment layers *(Röthlisberger and Lang, 1987; Fountain and Walder, 1998)* once reached the bedrock.

Throughout the melt season flow pathways are subjected a progression of their transport capacities as well as their position. It must be a seasonal variable progression of their development because of varieties in seasonal meteorological conditions and transformations of the ice surface. Even though there is assumed to be preferential flow-paths developing on similar positions as the year before. This theory can be underlined by the fact of preferential absolute positions of longitudinal and transverse crevasses.

The evolution of flow-paths is not completely investigated since there is no possibility to gain insight in the internal structure of the ice. Though there is the general consent investigated by closure rate and melt rate, that the pathway system must be an arboreal branching system *(Röthlisberger 1972, Shreve 1972*). It could be compared with the fluvial classification by Strahler. It is assumed that there must be downward increase of the englacial pathway average diameter and the capacity of water transport. The evolution of big englacial pathways is believed to proceed at the expense of smaller ones. Pore water and water channel pressure is the driving factor to enlarge the channel diameter.

The average dewatering velocities are considered to be significantly faster through the glacier conduit system than through the snow-firn layer. Notwithstanding, there is a huge range in transport velocities. *(Collins, 1995*) If we consider that the glacial conduit system is closed during the cold season due to freezing, it is obvious that such channels have to be widened in the beginning of the melt season to meet the transport capacity during the estival melt phase.

The whole glacial hydrological system is sort of a reservoir. It contains liquid water, which is stored to a lesser or greater extent in the glacial subsystems. The snow and firn aquifer is considered as the slower run-off reservoir, whereas the movement through the moulin-conduit system is remarkably faster. Isotopic and chemical meltwater investigations revealed a transit time of a few hours. *Behrens, Bergmann, Moser, Rauert, Strickler, Ambach, Eisner, Pessel, (1971)* describe the reservoirs with linear function properties (snow-firn aquifer), where dewatering is in- and decreasing almost linearly.

Over the ablation period the melt front is progressing towards the ice surface, where it gets locally collected and locally drained to the englacial conduit system. At the same time the melting front as well as the ice surface (once reached) built the water table. These are locally saturated zones or layers, where pores and spaces are completely filled with water. Dependent on overlying layers and water quantities pore water pressure is higher than in unsaturated snow and firn layers. The increased pore pressure leads to the effects described above.

With different approaches there has been tried to illuminate the complex englacial system. Big efforts were made to get behind the leaky system of the glacier body. Still it is not completely understood how and under which conditions and especially in which temporal context the cavity system is built. There is still a lack in understanding the close interaction between the actual processes.

The englacial system can build up very quickly depending on the various parameters. It is supposed that linked cavity systems of the anteriour years play an important role, as they can function as preferential paths for the present channels. But still old cavities and tunnels are not always “reused” in the following year. This is intensified by the fact that those conduits are closed and fully frozen in winter and early spring *(Stenborg, 1965; Lang, Schädler, Davidson, 1977).*

Other researchers do not divide this insight. Hooke (2005) insists that there are still open – non water-filled channels in the glacier body, if a step slope angle forces water to pass too fast to freeze in the conduit. Pressure in such open air-filled channels must be atmospheric *(Hooke, 2005*).

In any case the melted water from the snow cover in the beginning of the ablation season enters the englacial system causing the filling of the conduits. Due to the fact that not all of the conduits are opened and free to pass, water pressure in the conduits is significantly rising *(Collins, 1979)* In some cases it can be dammed up, which causes extreme pressures in the tunnel system. Glacier outbursts and sudden releases of the entrapped water can be the consequence *(Hooke, 2005, p.244).*
The so called efficiency of the tunnel system is lowest in the beginning of the ablation season *(Elliston, 1973; Lang, 1973).*
With the rising efficiency the drainage system is increasingly reacting faster to internal and external hydrological and meteorological influences. The reaction to those influences is greatest in the late ablation season, when storage capacity is lowest. The daily variation though appears to be more significant than the change over the whole melt season (*Collins, 1979)*.

According to data collected on South Cascade Glacier, Washington, USA water storage capacity maximum is reached when the melt front already propagated to the ice surface, but englacial conduits are still scarcely developed. Furthermore, it is suggested that at this time the conduits are still partly sealed off. Therefore outflow is impeded locally and leads to backwater effects *(Stenborg, 1970).*

During the ablation season channels and conduits are widened causing better dewatering. At the same time lower firn cover areas and greater bare ice areas result in smaller reservoir capacities finding its minimum in the late midsummer at the tail of the ablation season. Various investigations of glacier run-off on Gornergletscher, Switzerland, during the ablation seasons of 1970-1979 by Collins underline this alteration *(Collins, 1979).* Hydrograph recession analysis done after summery snowfall events resulting in negligible melt, do extract this information quite clearly. Moreover the method pointed out unambiguous the run-off background signals, originating from ground basal flow.

2.2.1.3 Storage and Movement on the Glacier Basis (subglacial water transport)

This ground basal flow is present over the whole season, but varieties in basal discharge amounts depend on the size and amount of the overlying ice. Even when there is no melt measurable small amounts of water are depleted on the shear zone of ice and ground. It is believed to be produced by geothermal heat and thermal energy generated by the basal sliding or the pressure produced by the weight of the ice. Studies pointed out that basal meltwater flow is not always entrapped in channels but flows in a thin layer of about 1 mm thickness *(Weertmann, 1972, as cited in Singh, 2001).*

This subglacial water film is retained as long as the pressure maintains on the pressure melting point *(Weertmann, 1972, as cited in Singh, 2001)*. If the basal melt flow exceeds certain amounts and pressures rates it will cause turbulent heating leading to the development of channels *(Hubbard and Nienow, 1997).*

There are coexisting theories how the basal water channel system is developing and especially how it can by physically described *(e.g. Shreve 1972; Hooke 1984).*

It is probable though, that processes steering the properties of subglacial channels can be very different and even of small scale and of local importance.

To *Hubbard and Nienow (1997),* there are to be listed “variations in meltwater flow, climatic and meteorologic conditions, overlying ice amount, surface slope, different sliding velocities, and substrate composition. Individual properties, especially regarding small glaciers are of great significance and can not be neglected. Furthermore there have to be considered special diurnal and annual differences of these interactions. One has to understand that glacial hydrology is not a sequence of dissipated processes, but a highly interacting hydrological system.

It is the reason why studies on sub- and englacial hydrology are hardly comparable. Quantitative information and qualitative understanding must be carefully extracted. Moreover this fact explains the existence of various theories and diverging results.

Nevertheless the basal flow induced by basal sliding or geothermal heat is only a background signal of lesser significance. Its variations and amounts are lesser than the superimposed depletion caused by daily melting. In order to get the melted water amount, background flow can be subtracted without great difficulties. The background flow is best visible in the early ablation season or in autumn, when melting is neglectable (best in wintertime, but most of the gauge stations for glacial run-off are not working).

Therefore the discharge at the glacier snout has different origins: (i) run-off due to rainfall, (ii) run-off due to melting of the wintry deposited snow cover, (iii) melting of the firn stratum underneath the snow layer or above the glacier ice and (iiii) melting of the glacier ice on the surface and on the contact zone to the ground.

2.2.2 Time Scale of Storage Processes

Glaciers can have storage properties retarding the run-off (in comparison with bare ground). Research on this natural reservoir and is of great importance as it impacts as well human settlements and civil use of the natural environment (e.g. hydroelectric power production). Storage in and on a glacier basically occurs within two different phases: liquid (water) and solid (ice, snow). Therefore any glacier represents a storage body for water if melted. This storage capability is differing regarding the different time-scales. There can be observed a huge variability of time in the potential retention of water, in regard to the different glacial zones. It rages from short-term water storage of only minutes up to long-term storage of hundreds or even thousands of years (**Fig. 2**) *(Jansson et al., 2003, p.116-117).*

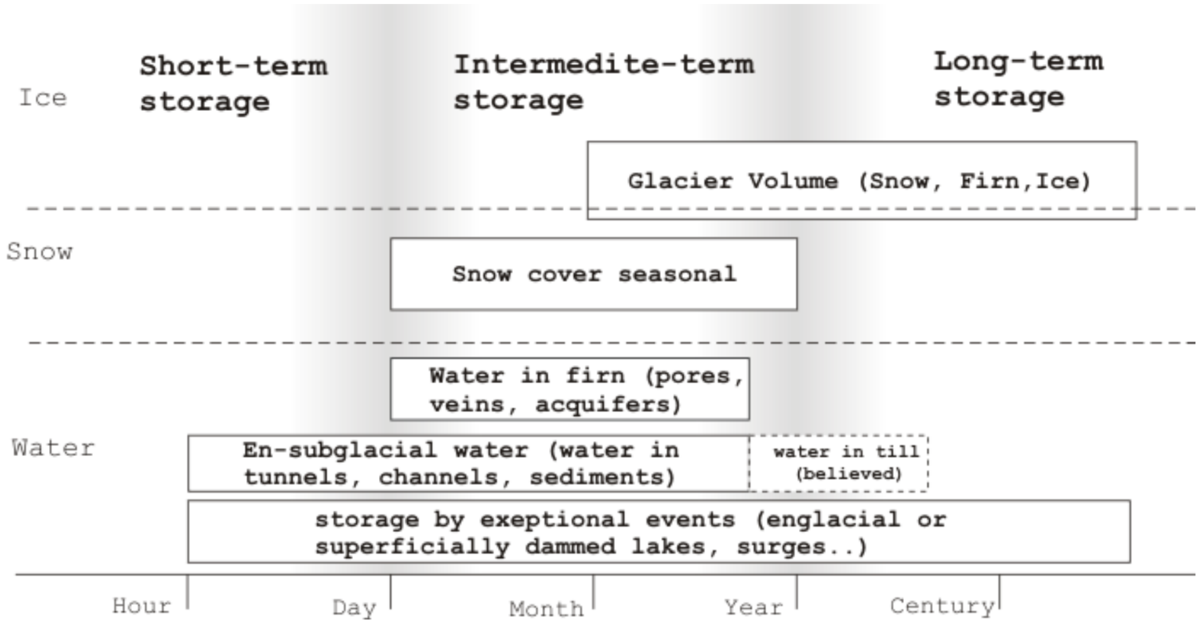


Fig. 2: Schematic figure representing timescales of water storage on glaciers (after Jansson et al., 2003)

2.2.2.1 Long-term Water Storage

Generally long-term water storage in glaciers is accomplished by the single ice body. The highest proportion of total water stored in glaciers lies in the ice volume itself. Only comparatively small amounts of snow and firn make out the total volume of a glacier. Nevertheless their importance for the watershed hydrology is not to be neglected since most of the ice body is not actively interacting during ablation seasons.

Glacier ice contains enormous amounts of frozen water. Worldwide 60% of the freshwater resources are stored in the world’s glaciers *(Hooke, 2005, p.3)*. Anyhow most of these freshwater resources lie in the world biggest ice sheets, the Antarctica and Greenland.

For temperate glaciers applies that despite their often comparable small aerial cover in a basin, they have a great influence in catchment hydrology.

Run-off variability is lowest at glacial cover percentages around 40%. An increase as well as a decrease in glacier cover leads to a higher variability in run-off *(Fountain and Walder, 1998; Röthlisberger and Lang, 1987).*

Run-off from not glacierized catchments is normally generated by precipitation and only partly by melting of snow. In completely glacierized areas the run-off is induced mostly by glacial melt. That means that glacierized catchments are dominated by energy induced run-off processes *(Chen and Ohmura, 1990; Lang, 1987).*

Long-term glacial storage is basically affected by the dominating climatic conditions. Glaciers grow only in an adequate climatic environment. Climatic variability is therefore the dominating factor for growing or shrinking of the total glacier mass. It is generally accepted that the velocity in which climatic shifts occur tends to be slower than the processes responsible for intermediate- and short-term storage *(Braun, Weber,Schulz, 2000).*

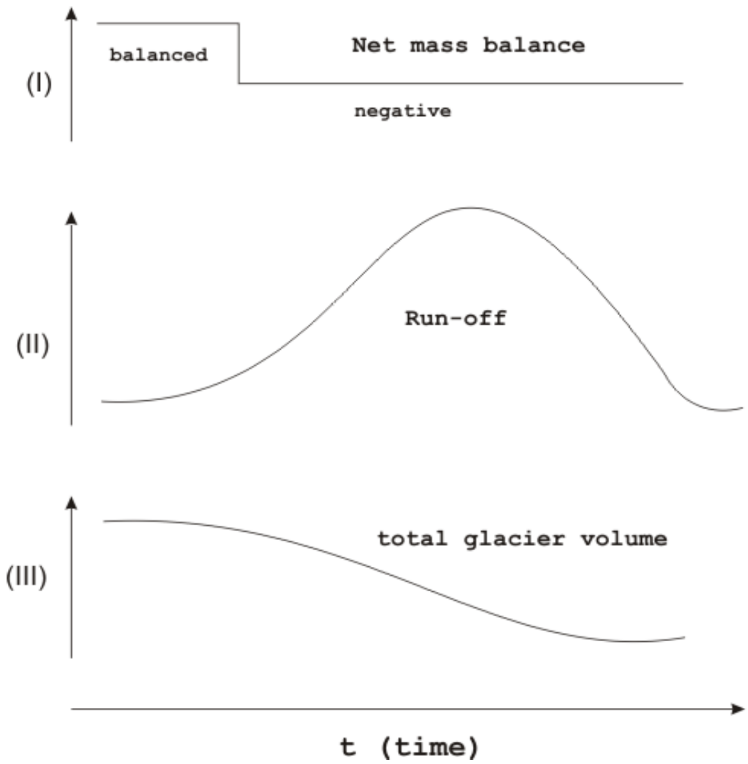


Fig. 3: (Schematic figure) Effect of negative mass balance (I) on glacier run-off (II) and total glacier volume (III). Run-off is responding with lag time to negative glacier budget and decreasing when total glacier volume is shrinking (virtually tot. glacier area) (after Jansson et al. 2003, p. 119)

2.2.2.2 Intermediate-term Water Storage

Intermediate-term storage of water refers mostly to water reacting in a time frame of about a day to a year. Huge amounts of water get released in summery melting. For temperate alpine glaciers this will happen mostly from June to October *(Escher-Vetter and Reinwarth, 1994; cited in Jansson et al., 2003; p.123).*

The intermediate-term storage of a glacier is the most complex in regard to the other time scales. Storage at this time scale occurs mostly because of the capability of firn to fill up the pore spaces with melt water and the retarded release of it, as well as the melt of snow as a proper storage of water itself *(Schneider, 2000).*

The wintery accumulated snow cover contains great amounts of frozen water. With rising temperatures and increasing solar radiation, melt processes are accelerated. Melt water percolates through the snow-firn body. Usually velocities of percolation are within a range of

0.1 to 0.3 m/h *(Schneider, 2000).* The percolating water can not easily be transmitted to the glacial drainage system. The ice is almost impermeable. Therefore access to the englacial drainage system is only possible at very local holes, entrances and crevasses. If the amounts melt water input exceed the drainage capacities it will lead to backwater effects and therefore to a storage of water in the different glacier aquifers. With the progression of the melt season discharge conditions are becoming more effective resulting in reduced storage *(Jansson et al., 2003, p. 122).* Retended melt water at the beginning of the ablation season can be released later on in summer.
Even if often observed, wintery storage of liquid water is hardly investigated, because of the impeded field conditions.

Summery rainstorm events let run-off respond quickly. With the intermission of rainfall, run-off is decreasing in a certain rate, depending on the storage characteristics of the glacier and especially on the period of occurrence in the annual cycle. Calculation for Storglaciären and South Cascade glacier showed normal pre-storm flow rates were achieved not till 1-2 weeks after the event *(Östling and Hooke 1986, cited in Schneider 2003, p.121). Behrens et al. (1982)* indicates a timeframe of about several days to weeks for the Vernagtferner. Due to capillary forces pores in the firn can release the water only slowly and retard the run-off. Depending on the channel development in the firn this process can be significantly vary.
On the contrary bare ice is directing water relatively fast to the outlet. The measured time window between melt and run-off is about only a few hours *(Nienow, Sharp, Willis, 1996).*
Moreover till sediment layers at the glacier bed can store water as well. They can attribute to the wintery run-off and decelerate the reaction of run-off to melt- or stormwater input. It is believed that the measurable up-lift of the glacier at the beginning melt season is linked to the annual intermediate-term water storage and the till layer. Water pressure in and under the glacier is rising. As a consequence the glacier swells and its surface is lifted measurably *(Iken et al., 1983; cited in Jansson et al., 2003, p. 121).*

2.2.2.3 Short-term Water Storage

Short-term water storage considers diurnal variations of run-off. Storage processes on a short-term timescale have been investigated in several studies *(e.g. Collins, 1979; Hock et al., 1999; Schneider, 2000 and others).*

Those variations can have different origin, but are basically caused by the diurnal temperature and radiation cycle. Melt is reacting immediately to rising temperatures and radiation. Nevertheless the characteristics of the reservoirs influence the velocity of discharge reaction *(Oerter, Baker, Moser, Reinwarth, 1982).* E.g. *Hock and Nötzli (1997)* identified three different reservoir types for aerial melt and discharge modelling at Storglaciären. Hydraulic conductivity values (k-value) were assigned to the different reservoirs. The highest values where assigned to the firn reservoir and the lowest to the ice reservoir. The snow reservoir has its greatest impact on the discharge at the beginning of the melt season. Therefore progression of the melt season spotlights the different reservoirs. The retention of water through reservoirs must be allocated time-dependent. In other words the amplitude of short-term run-off is highly dependent on the annual period, literally on the dominant reservoir at this time of the year. The up-wandering snowline points out the switch to another reservoir *(Collins, 1998).* The transition of the snowline is also accompanied by the transformation of the subglacial and englacial drainage system *(Seanberg, Hooke, Wiberg, 1998)* and the intersection of a distributed run-off regime to a channelized run-off regime *(Nienow, Sharp, Willis, 1998).* The accumulation to ablation area ratio articulates this transition in one term.

Additionally to the discharge caused by infiltrating melt water run-off can be simply provoked by liquid precipitation. It is generally accepted that melt processes in snow or firn are accelerated by the sensible heat in rain and the formation of preferential flow tracks.
This means that the measured total run-off can exceed the total fallen precipitation in a rainstorm event *(Singh et al. 1997).*

3 Study Region and Investigation Area

3.1 Hohe Tauern and the Goldberggruppe

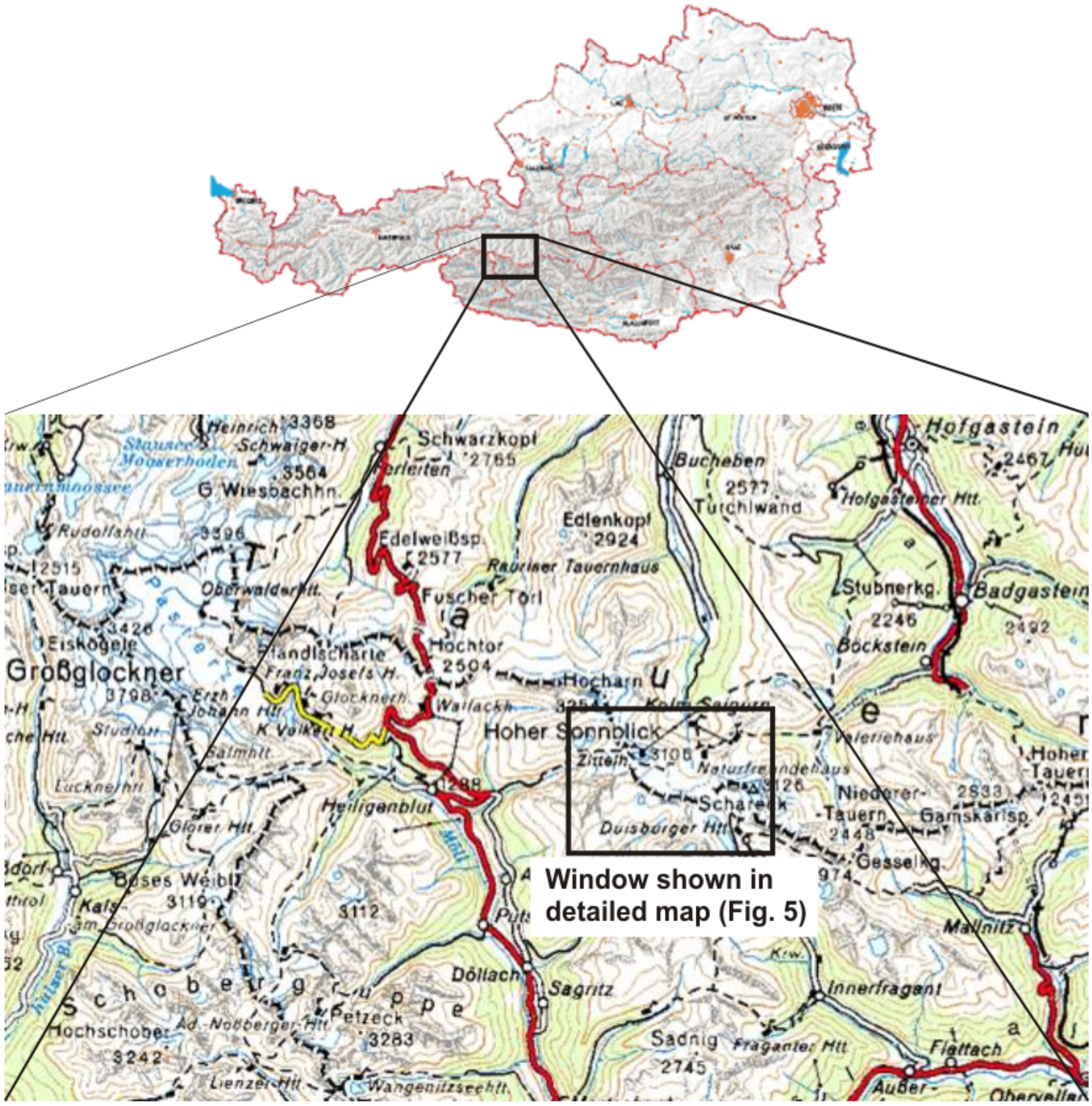


Fig. 4: Topographic overview Austria; Overview of Sonnblick Group with Großglockner, scale 1:500.000; Austrian Map online ÖK500, access April 2008

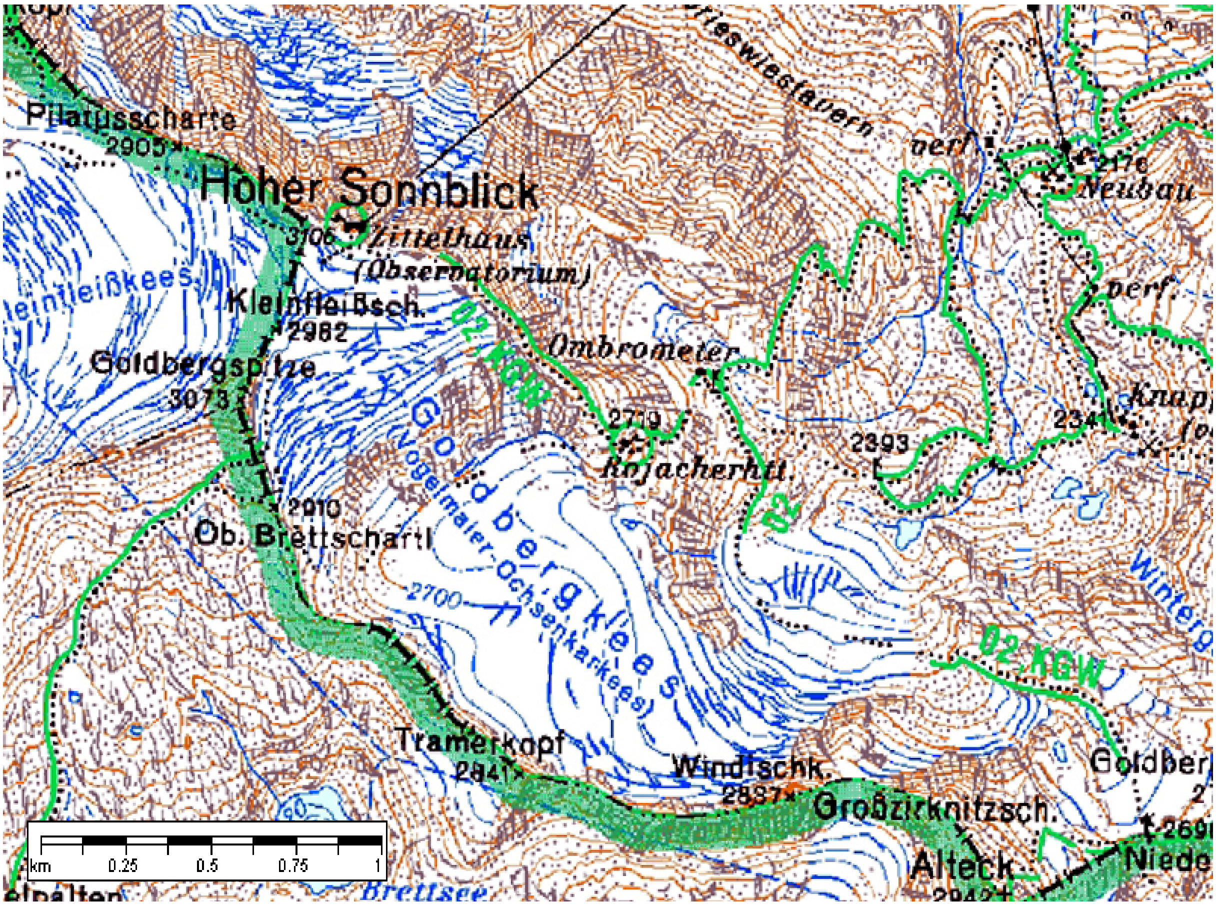


Fig. 5: *Hoher Sonnblick* with *Goldbergkees* and Kleinfleißkees; scale 1:50.000, (Austrian map online, access: April 2008)

The Investigation area is situated in the Austrian Eastern Alps in the so called Hohe Tauern. The Hohe Tauern is part of the main ridge of the Alps, built by the highest elevations. The Hohe Tauern ridge is more than 140 km long and runs in West-East direction. At the same time it is the centre of the Eastern Alps. The Goldberggruppe is a subdivision of the Hohe Tauern and lies in the eastern part of the Hohe Tauern.

The Hohe Tauern is a climatologically important mountain ridge in the Eastern Alps. They represent a climate divide. The Hohe Tauern are characterized by a number of 101 small and middle glaciers with a total area of about 90 km². The glacier border line lies between 2700 and 2900 meters. (Groß, 1987). The size though is shrinking fast since the last glacier advance in 1850 and accelerated even more in the last decades.

The Hohe Tauern counts a great number peaks with elevations over 3000 meters. The highest peak of the Hohe Tauern and at the same time of Austria is the Großglockner with an altitude of 3798 meters. The Hohe Tauern lie in the three Austrian regions Salzburg, Kärnten and Tirol. A smaller part is bordering on the Italian frontier.

Most of the area is accounted as national park with a total area of 1836 km². The nature protection area is subdivided in a core zone and in buffer zone (peripheral zone).
Geologically, the Hohe Tauern are composed of morphic bedrock and shist. Geologists term the special formation as the “Tauern Fenster”. The characteristics of the rocks built special geomorphologic forms and delineate the distinctiveness of this alpine region.

Economically, the region is mostly dependent on tourism and on local low rank economics. Important though is the power industry which runs some important storage power stations. The reservoirs are veritably linked to the summery incidence of glacial meltwater.

Therefore it is also a crucial area regarding meteorologic conditions and precipitation.
The Goldbergruppe is a subdivision of the Hohe Tauern and lies in the two regions Salzburg and Kärnten. The highest elevation is the Hocharn (3.254 m). Another famous peak is the *Hoher Sonnblick* with 3106m of altitude. The Goldberggruppe is located in the Eastern part of the Hohe Tauern, only about 20 Kilometers from the Großglockner (3798 m).
The onomastic of the mountains comes from the mineral richness of the region. It was a known mining region for gold. Nowadays mining has been dismissed. Only some mining relicts like drifts and old barracks can be found.
The Goldbergruppe houses some small sized glaciers like the Hocharnkees, Krummlkees, the Wurtenkees, the Kleinfleißkees, the Wurtenkees, the Schlapperebenkees and the *Goldbergkees*.

3.2 The Goldbergkees

The *Goldbergkees* has a total area of about 1,42 km² (measured in 2003). It is part of the Sonnblickgroup in the Hohe Tauern. When compared with other glaciers it is only of small size. Its influence concerning hydrology or landscape is therefore only of local interest. Though, it is one of the best investigated glacierized areas in the Alps. It can be attributed to the special position of the glacier itself. It lies on the northern border of the main ridge of the Alps. The meteorologic circumstances are varying strongly and depend on the current weather conditions. The nearness to a continuously operated meteorologic observation station (since 1886) at the *Hoher Sonnblick* (3105 m) provides reliable data as groundwork for any further investigation. Moreover the size of the glacier makes it sensible to any climatic shifts and makes apparent changes e.g. in mass balance, size, length, thickness or geomorphologic and hydrological processes linked to the glacier ice.

For this reason studies on the *Goldbergkees* can easily be adapted to similar glacierized zones in the world but especially to the east alpine glaciers.

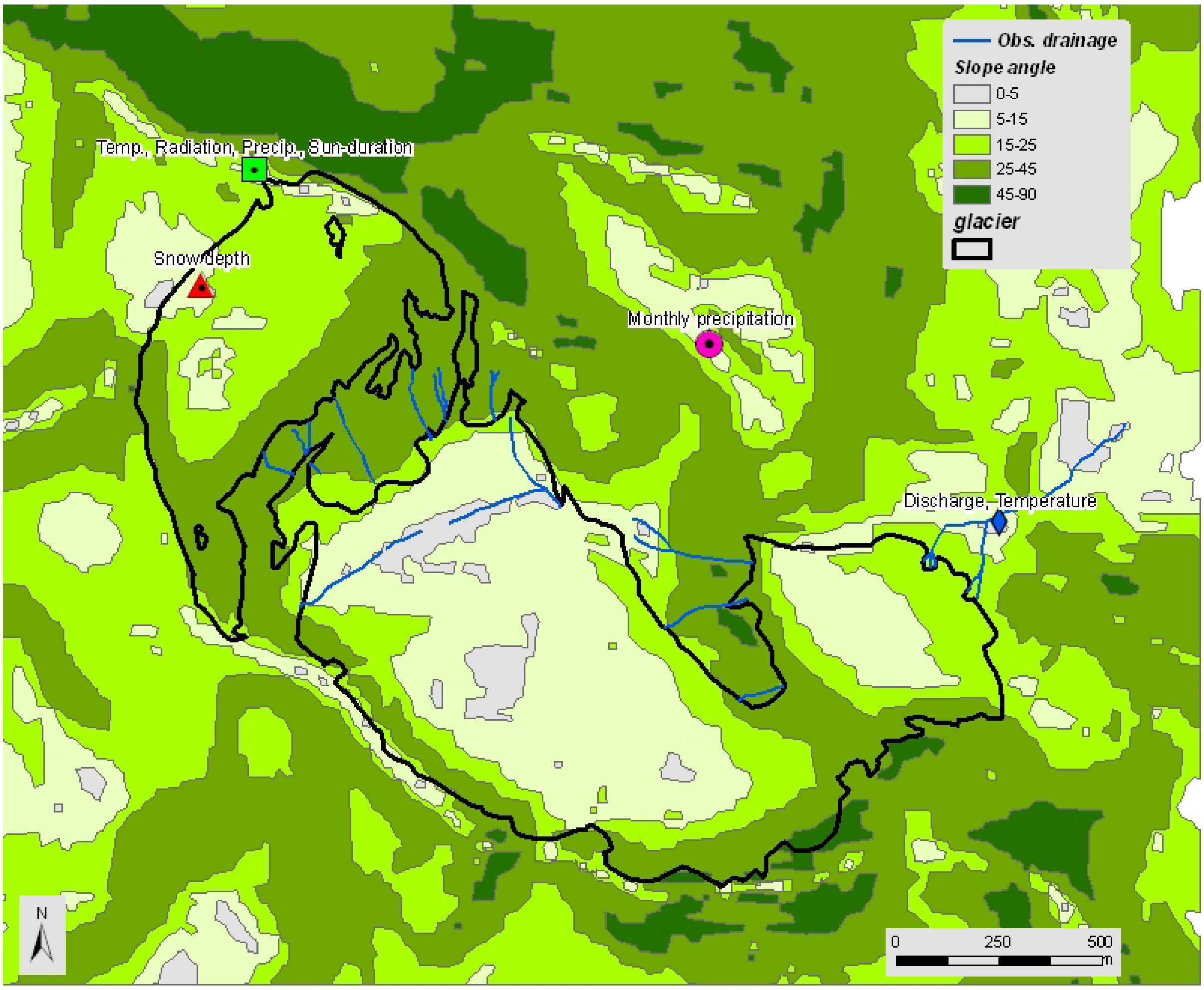


Fig. 6: *Goldbergkees* Slope angle (in degree) and position of the hydro-meteorologic stations in the area. Glacier extension digitized after aerial image 2003

3.2.1 Glacier

The glacier has a total area of about 1,42 km². It can be subdivided in „*Oberster Boden*", „Oberer *Boden*", and „Zunge" (tongue). The „Oberste *Boden*" is the highest part, situated directly underneath the *Hoher Sonnblick* peak. It has an area of about 0,39 km² and is separated through a ripe rock face of the "Obere *Boden*". The "Obere *Boden*" is the main part of the glacier. Together with the tongue it has an area of aproximately 1,01 km² (tongue has approximately ¼ of the area).

3.2.2 Climate

The climate on the *Goldbergkees* can be classified as a high mountain climate.
One of the main climate components is temperature. For the climatic period of 1961 to 1990, typical characteristics of the seasonal cycle is a minimum in temperature in February and a maximum in late July and beginning of August *(Auer, Böhm, Leymüller, Schöner, 2002*).
Radiation is highest in the ablation season in summer. Radiation follows the annual radiation cycle. Nevertheless there can be made out great differences from year to year. Solar radiation is dependent on the sky cover. In years with increased cloudiness radiation is reduced significantly.
Depending on the temperature precipitation is falling in solid or liquid form. Precipitation in winter is accounted for snowfall whereas rain is dominating in summer. Notwithstanding depending on actual temperatures and altitude also snow is a common phenomenon in the melt season. Over 80% of the annual precipitation quantities are falling as snow. The summerly period accounts for the greatest precipitation amount of over 1300 mm (1961-1990), whereas mean wintery deposition is around 800-900 mm. The deposition rates are strongly dependent on local criterias as exposure and orientation (Luv or Lee slopes), *(Auer, Böhm, Leymüller, Schöner, 2002*). Summerly snowfall influences greatly the melt and hydrological processes on the glacier surface. In transitional periods as spring and autumn, rain instead of snowfall and rain can alter, depending on the temperature. Rainfall on snow causes accelerated metamorphosis processes and changed hydrologic conditions. Snow can soak the rain and therefore refreeze in the snow layers. Wintery snow heights are depending on the elevation, even if the elevation increasing factor considering precipitation is lower than suspected.
Snow heights reach their maximum in late spring or at the beginning ablation season, much later than in lower situated areas. The peak in snow height is temporally delayed. Snow heights are varying enormously. Wind drift before and after deposition, as well as

avalanches are redistributing snow and depositing it on preferential areas. They can vary from about 1 Meter to 7 Meters and more for the same year. In water equivalents snow height vary from under 1000 mm for wind exposed areas to about 2700 mm for wind exposed zones*. (Hynek, 2000; Mott, 2006).*

Usually melt is not taking place before May. Increased run-off from melt is initiated when the snowpack is isothermal. To simplify comparability, the ablation season for the *Goldbergkees* begins with May and ends with September.

Wind its transport capability for snow is an important factor in winter. On the *Goldbergkees* the most frequent wind direction is South-West. A second frequency maximum is North-West. Depending on the weather situations wind directions are highly rotative. Because of its orography, slight shifts in general wind directions can cause huge changes in wind fields over the glacier. The maximums for wind speed can be accounted for January. Generally the winter season shows higher wind speeds in average and for wind squalls*. (Mott, 2006)*

The total ice volume of the *Goldbergkees* is more than 48 million m³. Nevertheless it decreased since the last glacial high stand in 1870 for more than ¾ (**Fig. 7**).

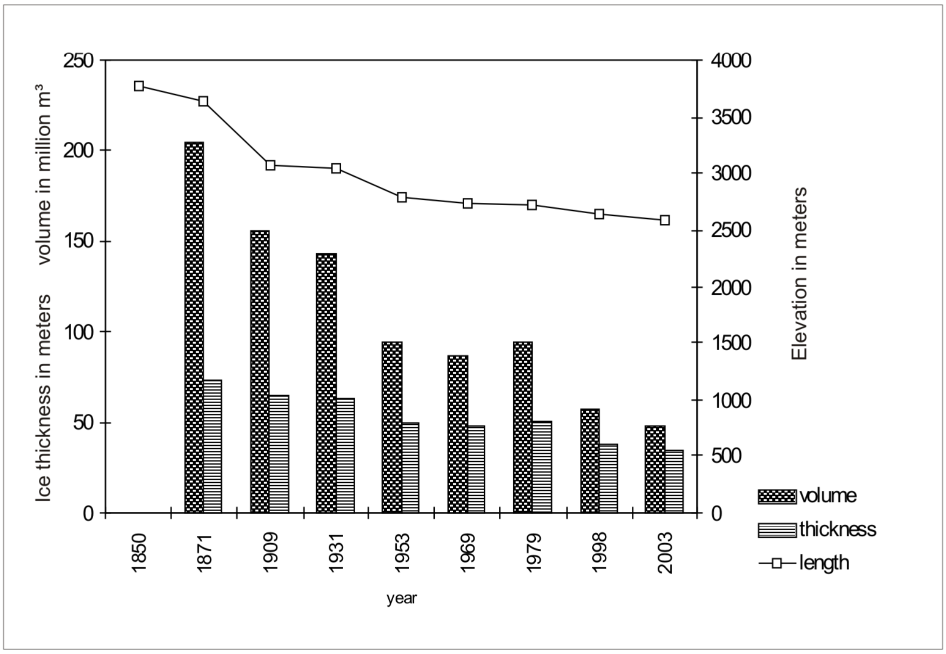


Fig. 7: Volume, thickness and length of the *Goldbergkees* for the selected years from 1850-2003 in absolute values (Böhm et al., 2007)

A crucial aspect of glacial retreat is the fact that length variations do not react with the same magnitude as volume variations. When volume is affected by snowfall or melt, length changes are responding slower or with a retarding factor. Melting affects more thickness and therefore volume changes as for example length changes. This may be arise from the special shape of this glacier, but could in most cases be adopted for other similar sized alpine glaciers as well (**Fig. 8**).

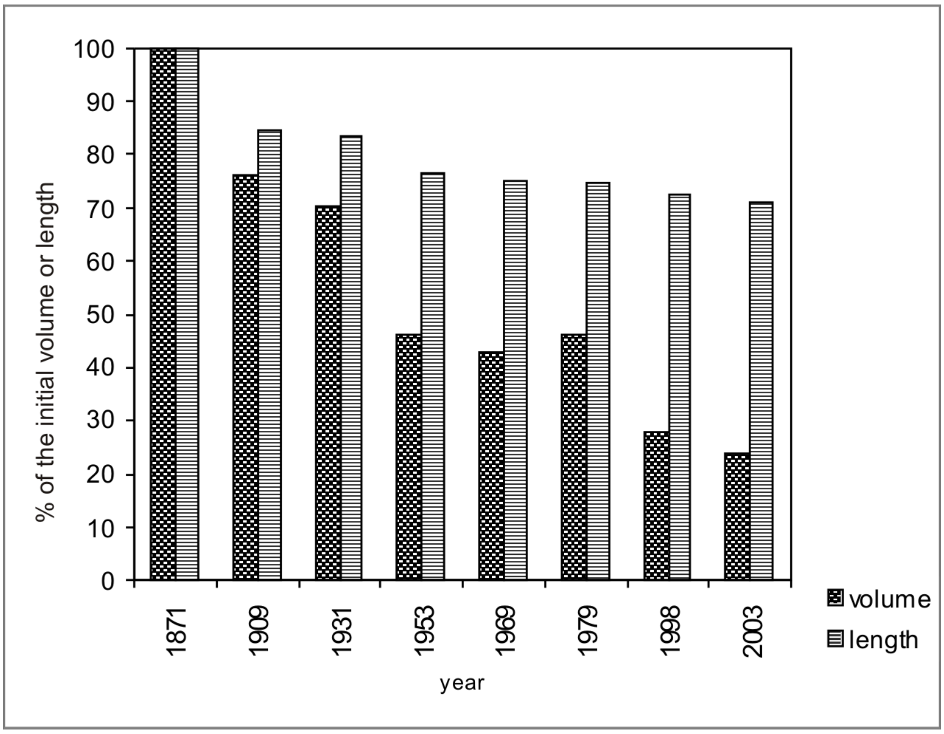


Fig. 8: Volume, thickness in relation to value in the year 1871 in % (Böhm et al., 2007)

4 Methodology

4.1 Tracer Studies

Tracer hydrology is a well-established method in glacial hydrology, but originated from karst hydrology. Since the late 19th century salt and dye tracer investigations are known methods for investigating glacial drainage systems. Normal salt or natural or chemical dyes are injected to measure flow velocities as well as amounts of discharge. In some cases it can provide also conclusions of geological, meteorological, biological or chemical character. Therefore it has been often used and provided important information on glacial hydrology.
The injection site is situated upstream and represents normally a fastly draining moulin or crevasse or simply a supraglacial melt channel or gully. The detection site is normally situated on the glacier mouth or further downstream.

Initially salt injections were made but reached their limits relatively soon as it came to longer flow distances or greater water quantities. The detection limit of salt in water is only in the mg/l range. It turned out that other substances had much lower detection limits. Nowadays Fluorescein and Rhodamine are the most commonly used chemical fluorescent tracers in glacial hydrology. Their detection limit lies around 0.0006 ppb (part per billion). It can simply be detected by manual sampling or by an in-situ fluorometer optionally with a (automated) data logger. Ideally, the results should point out rather clear tracer breakthrough curves. *(Fountain, 1993; Sharp et al., 1993).*

4.1.1 Tracer Measurement and Recovery

For a precise investigation on the characteristics of a glacial hydrologic system tracer recovery plays an important role. By knowing the amount of injected dye tracer and the total run-off on the glacier mouth we can calculated the quantitative percentage of dye bypassed the fluorometer. This can be a meaningful approach for investigating storage characteristics. A loss of dye tracer points to storage processes inside the drainage system. Nevertheless the breakthrough curve can flat out so much that a distinction from the background signal is no longer possible. Water and in this case dye tracer is hardly stored permanently. Often the measurement period is too short to get the whole amount of initially injected dye tracer. Roughly speaking the greater the injected dye amount the better the percentage of recovery (in case the measurement period is sufficient and maximum detection limits are not overstepped).

High percentage recovery rates indicate a fast and well developed drainage system in which storage is only of short term. According to various studies; efficient or well developed drainage systems show transit velocities of more than 0.2 m/s *(Moser and Ambach, 1977; Lang et al., 1979; Burkimsher, 1983 in Hubbard et al. 1997; p. 942).*
Low percentages of dye recovery points to an arborescent, “delicate” and/or hardly developed drainage system. The storage can be intermediate (hours) or long term (days, months) *(Behrens et al., 1971; p.98).*
Tracer recovery is also influenced by the chemical and physical composition the melt water. Suspended sediments generally occur in subglacial channels can absorb dyes significantly. In consequence tracer recovery is affected. It depends highly on the dye. Some tracers are easily influenced by adsorbation processes whereas others remain relatively stable.

A second parameter diminishing tracer recovery is turbidity. Turbidity induced by suspended sediments and organic matter dislocates the fluorescent wavelength peak. Especially flushed-in organic matter does have similar wavelengths as certain tracers. Investigations employing Eosin and Fluorescein for example are highly interfered by soluted organic matter. The percentage of tracer recovery is therefore often reduced significantly. If turbidity is high enough and tracer concentrations relatively low, results can be difficult to interpret.
Normal turbidity rates though can be easily substracted from the tracer breakthrough curve easily. Some fluorometers can recalculate the effective tracer concentration in situ. *(Smart and Laidlaw, 1977).*

A third parameter influencing tracer recovery is radiation. Whereas some tracers are relatively insensitive to solar radiation others react quickly to daylight exposure. The choice for the appropriate tracer in field studies is affected considering the physical and chemical decomposition due to radiation affect. If solar radiation amounts are known exactly it is possible to recalculate the concentration. By having decomposition matrices composed of incoming radiation quantities and time of exposure it is possible to sum up to the percentage of tracer recovery possibly measured without exposure to daylight.
For the prementioned problems hydrological interpretation through tracer recovery rates is rather complex and can lead to a serious of false conclusions. Therefore it is of utmost importance to select the involved tracers carefully and avoid any further sources of errors in the field. The arising problems should be concerned already in the run-up to the field investigation *(Nienow, 1993).*

4.1.2 Dye Tracers

4.1.2.1 Coloured Dyes

Non-fluorescent coloured dyes are widespread in hydrological as well as soil studies. They work on other principles than fluorescent dyes. The detection is rather qualitative than quantitative. They work rather on visibility than on measureability. Moreover they are easily adsorbed by chemical processes in the aquifer. Nevertheless they can provide important initial information about water movement in and on glaciers. In field or laboratory they can be measured with a simple spectrometer. The spectrometer measures the extinction of light for determined wavelengths. Similar to the fluorometer for fluorometric dyes, concentrations can be obtained by knowing the extinction curve of the dye and the actual extinction in the measured solution. The detection limit though is high, compared to their fluorescent relatives. Disadvantages are therefore greater injection quantities or shorter measurement distances.

In this field study Potassium permanganate ($KMnO_4$) was used to determine dispersion of water in snow and the propagation of the melting front in early summer glacial snowpacks.

Potassium permanganate is often used in soil sciences for pigmenting soil profiles. Potassium permanganate is an anorgnic salt of pink colour. It is water-soluble (64.0 g/l at 20°C). It is better visible on snow than other coloured tracers. Bluish, greenish and yellowish colours are not easily detectable in water.

4.1.2.2 Fluorescent Dyes

Fluorescent dyes are of common use in hydrology. In the past they have been used for various investigations in the field of glacial hydrology, but mostly for time discharge relationship measurements in sub- and englacial drainage systems.

The most commonly employed dye tracers are

1. Fluorescein
2. Sulfo-Rhodamine (G,B,WT)
3. Eosin
4. Phyranine
5. Lissamine
6. Tinopal

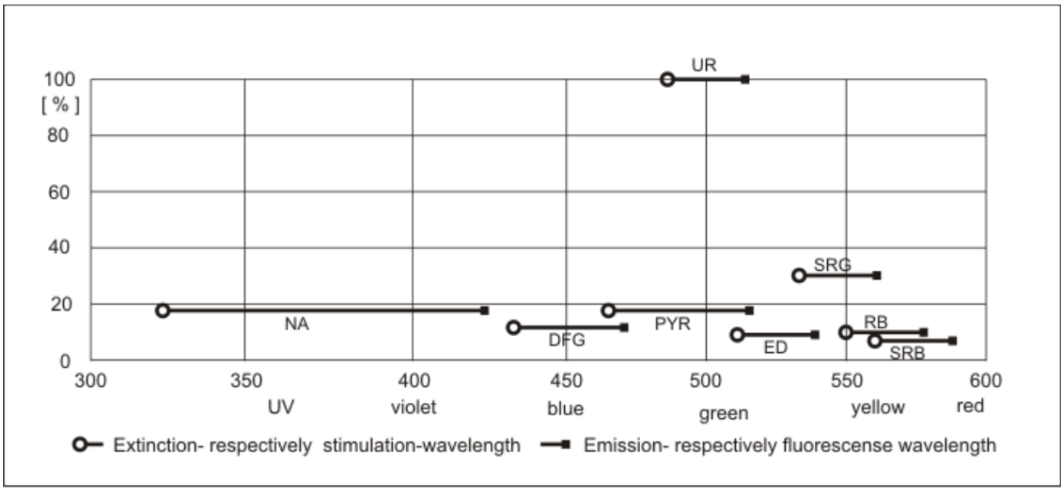
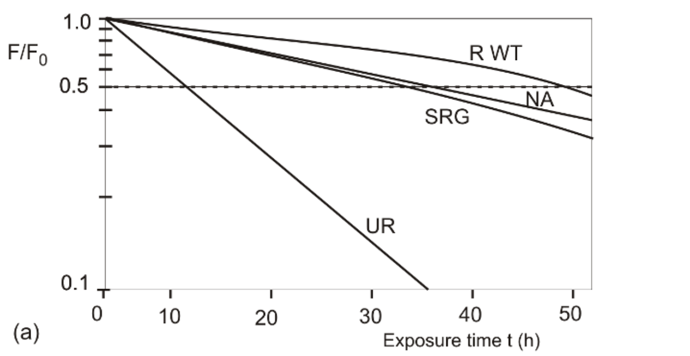


Fig. 9: Extinction and fluorescence wavelength of various fluorescent dye tracers; *Source: Wernli, 2003*

Fluorescent tracers are classified in the group: Artificial Dyes. That means they are not of natural origin. Nevertheless they are organic compounds. Stimulated by light they emit visible light with a certain wavelength which can be recorded by an automated field-fluorometer or analysed in the laboratory. The light wavelength after stimulation is approximately 300-650 nm.

Not all of the fluorescent dyes are adaptable for glacial hydrological studies. A predominant number is only poorly water soluble. As a consequence its application must be carefully elaborated. Dyes with good characteristics for hydrological purposes are oftenly applicated. The advantage of fluorescent tracers compared with coloured dyes, are their low detection minimas. For field studies weight and transport are important, especially for glacier investigations. Low detection limits need minor amounts of injected tracer. Their affortableness is of benefit as well. Furthermore they are toxicologically harmless. Nevertheless they show enormous differences in their optical, chemical and physical characteristics.



Fluorescent dye	T 1/2 Dye / T 1/2 Fluorescein
Eosine	0.6
Fluorescein	1.0
Pyranine	4.3
Amidorhodamine G	50
Rhodmine FB	69
Sulforhodamine B	71
Sulfoflavine	73
Rhodamine WT	93
Naphtionate	3.7

(b)

Fig. 10: Logaritmic Decomposition of fluorescent dyes exposed to radiation (a), and half life period in comparison with half-period value of Fluorescein (b); *Source: Wernli, 2003*

Fluorescent dyes emit radiation if stimulated, but lose their emission capability rather quickly. If exposed to sunlight they are decomposed and concentrations seem to be lower when reaching the detection fluorometer. The decomposed molecules are no longer capable of emitting light. The decomposition of a certain amount of fluorescent dye equals strongly the half value periods of radioactive substances. A fast decline at the beginning is followed by a logarithmic decline of the fluorometric potential.

Fluorescent dyes can be adsorbed by soluted sediments as well as by organic matter. The adsorbtivity of dyes and the influence of organic matter to the light emission record are widely scattered. Furthermore fluorometric emission is depending on the temperature and on the acidity (pH-value).

4.2 Field Study and Tracer Experiment Sequence

The following chapter emphasize on the steps undertaken to achieve the desired aims.

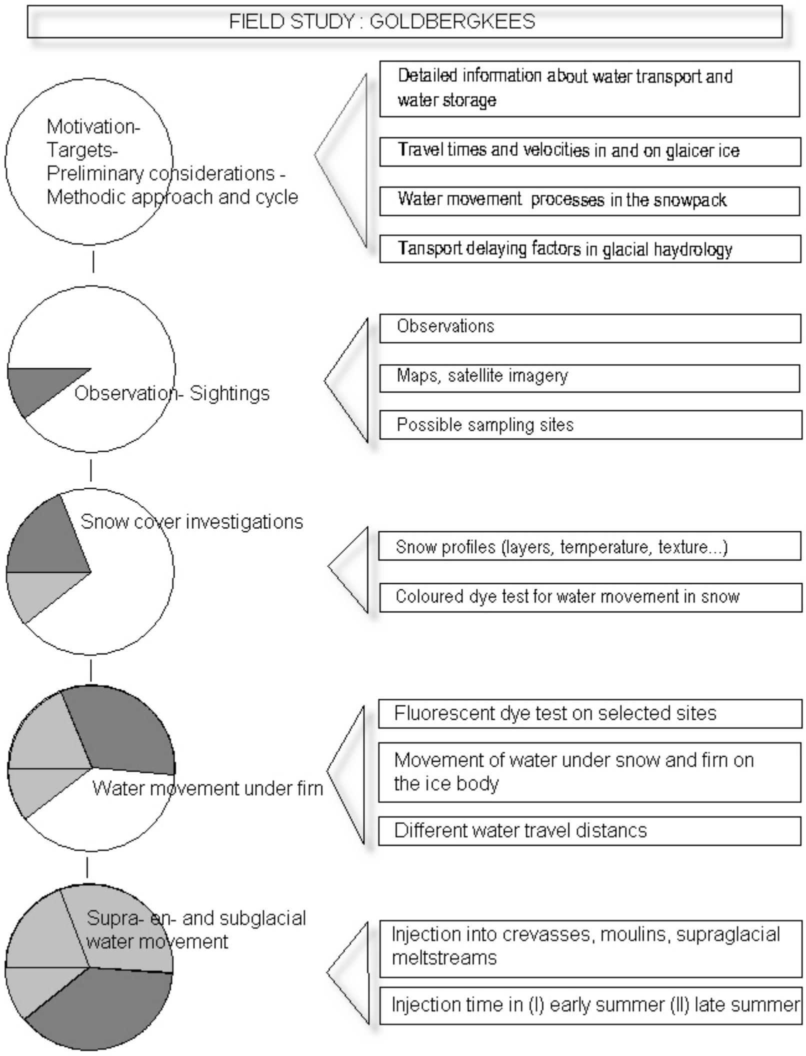


Fig. 11: Study sequence, time efforts for the field study on *Goldbergkees*; Source: own illustration

4.2.1 Pilot survey for experimental setup

Due to previous universitary excursions and field trainings of glacial mass balance measurements it was possible to get to know the area previous to the field investigation. In 2005 and 2006 I traveled with the Department of Meteorology as well as the Department of Geography and Regional Research (University of Vienna) to *Hoher*

Sonnblick Group. The idea for the current field study evolved in the lacking information about flow processes on the *Goldbergkees*.
In order to draft a fundamental frame of reasonable experiments a first aerial sighting was crucial. With which methods is it possible to gather information of flow and storage processes? Which information could be derived out of aerial monitoring?
Sampling sites where preselected on the basis of photographic documentation as well as observation of preferential supraglacial meltstreams, moulins and crevasses. Further information was extracted through cartographic means (Geo-information, maps, aerial images). The result was a table and a map (sketch) with collected potential sample sites for different investigation steps. The outcome of this is a flow chart for further field experiments (**Fig. 11**).

4.2.2 Snowpack investigations with the coloured dye Potassium Permanganate

4.2.2.1 Aerial dispersion of coloured dye on the snow surface

The seasonal snow cover is partly transformed to glacier ice in the annual cycle. Most of the snowmass on temperate alpine glaciers with almost no tongue movement, melts in the summerly ablation period. The target of sampling the snow cover is to trace the movement of water in snow. At the beginning of the melt season the snowpack is still partly frozen. Snow profiles determine exact texture, temperature and thicknesses of various snow layers. Potassium Permanganate as a coloured dye was used to colour the meltwater. On the one hand Potassium Permanganate is injected in different depths of the snow cover, ideally on the surface, in the top layer of the snowpack, over and under ice layers as well as on the firn or ice surface. The measurement was taking place from the 12th to the 14th of June.

The first steps were to choose the right sampling sites according to criterias like inclination, undisturbed snow layers, representability of snow layer composition (avoiding sites with wind drift, crevasses, shading).

On the chosen experiment sites were done snow profiles to record the snow layer composition as well as the current temperature and humidity distribution. In this case two profiles for three experiment sites were digged (two sample sites were of similar characteristics and minor distance, so could be described by only one profile).

Potassium Permanganate was disolved in water and dispersed with a handspray on a quadratic area of about 1 sqm. Due to the magenta colour of soluted potassium permanganate, the albedo would have been reduced significantly. Therefore the sprayed area was covered cautiously by a 5 cm layer of snow.

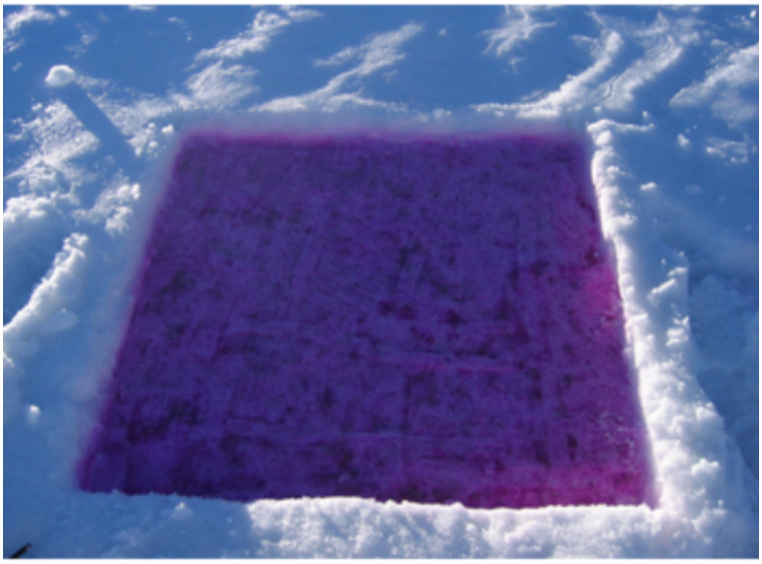


Fig. 12: Potassium Permanganate application on snow for permeability test; Source: own picture

It is important to not exagerate the spread water amount in order to not accelerate the infiltration into the snow body due to additional water. The process was repeated on two different sites in the morning hours (at 9 a.m.) (For detailed positions go to **Fig. 8**). After 17 hours a longitudinal cut was digged to document the propagation of the meltwater. Documentation was done by measuring distances and depths and recorded manually as well as photographically. After 37 hours from the dispersion moment a second longitudinal cut was digged to document further propagation. The longitudinal sections were build following the natural inclination of the terrain, beginning in the left upper angle of the coloured square, then after 20 h on the opposite angle of the square. For this reason two profiles were extracted out of one sample site.

4.2.2.2 Punctual Injection of Dye into the Snowpack

In a second experiment snow was punctured with small amounts of Potassium Permanganate, injected directly in certain depths into the snowpack. For a clear record after the excavation it was important to inject the Potassium Permanganate solution exactly in the right depth, without contaminating other snow layers.

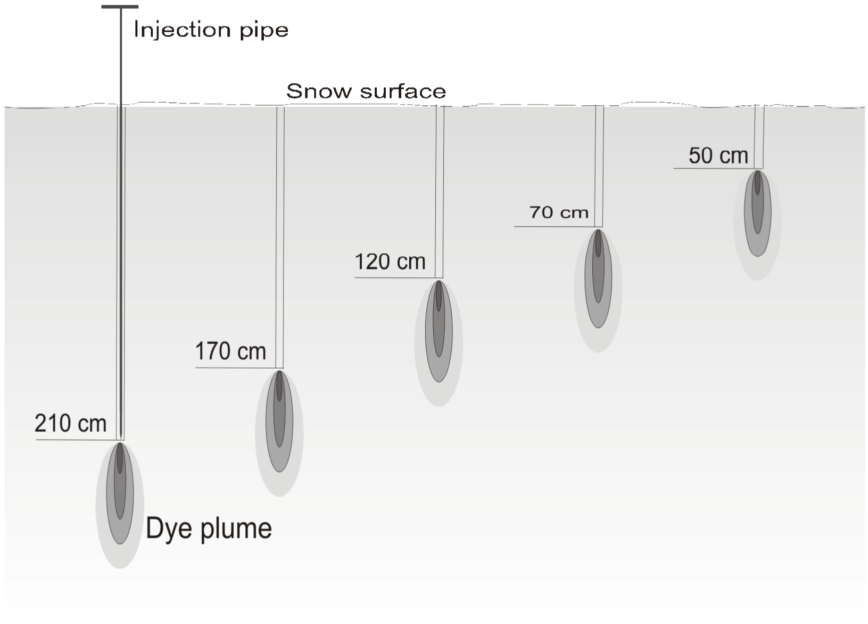


Fig. 13: Schematic view of dye injection into different snow depths with injection pipe. *Source: own illustration*

For this reason the Department of Materials of the University for Soil Sciences (BOKU) and Dr. Gernot Koboltschnig of the Institute of Water Management, Hydrology and Hydraulic Engineering help me design a special injection device. It had to be long enough to reach deep into the snowpack, stable enough to penetrate the hard layers and should hold a certain amount of fluid. Moreover the fluid outlet when released had to be exact.

Therefore an avalanche sonde of 4m length was adapted to serve as gigantic syringe or pipette. The syringe could hold up to 30 ml (with subunits of 5 ml) of fluid and could be adapted to any length if necessary.

With the aid of this injection pipe 30 ml of a saturated Potassium Permanganate solution (equal to 1.95 g dried matter) was injected side by side into the snowpack. After 24 hours the profile was excavated to document the development.

Meteorologic conditions as well as the physical characteristics of the snow layers with the observed flow and percolation paths of meltwater have been related.

4.2.3 Measurement of Water Movement on Firn

When snow melts on the surface mostly due to solar radiation and sensible heat fluxes water is mediated through the snow body to the subjacent firn. Glaciologically, firn is defined as snow endured a total ablation period *(Fountain and Walder, 1998; p.324).* firn is denser than snow because of decomposition of the hexagonal structure of the snow grains, refreezing and compaction. It can not mediate meltwater easily through its body. Therefore water flows primarily on top of the firn layer.

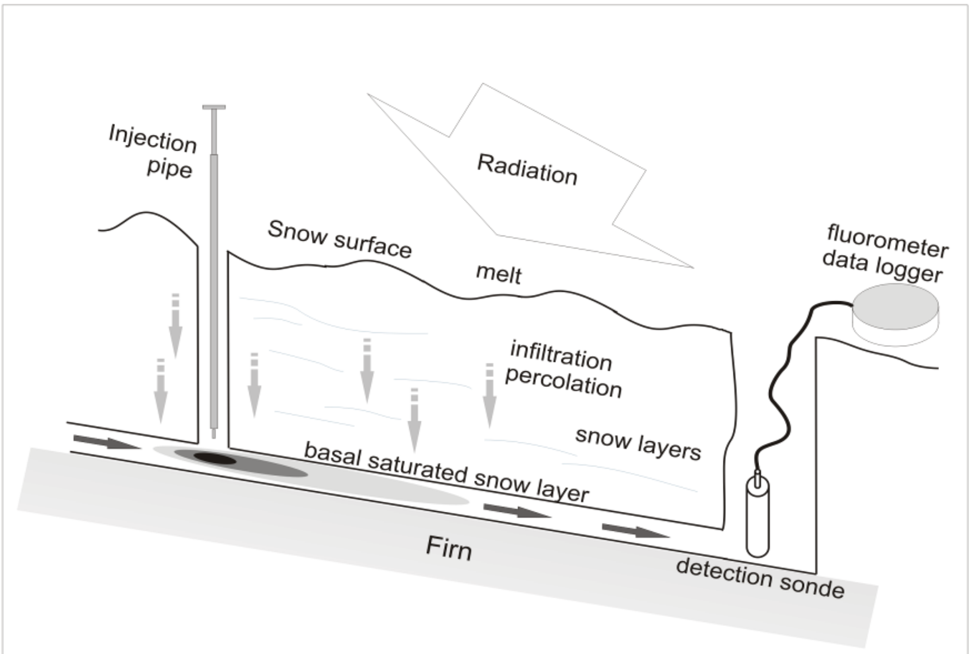


Fig. 14: Schematic figure; experimental configuration for dye tracer measurements on firn. *Source: own illustration*

For a coarse quantification of the flow velocities of water on firn, fluorescent dye tracer test were done. Injection of Sulfo-Rhodamine G as well as Naphtionate by an injection pipe and the following detection through the field-fluorometer generated a rough estimation of transit times (**Fig. 14**). The concept for the experiment is mostly adopted from Campbell, 2006, p. 969-985. The experiments were carried out from the 12th-14th of June.

The measurement could only be carried out where crevasses or moulins do not disturb the continuous water flow. In order to detect any tracer the fluorometer had to be installed in the direct slope line of the injection site. Nevertheless detection could be difficult if the supra-firn meltwater flow is channelized. If the flow paths do not match with the detection site no results could be obtained. Higher discharge causes more canalizing. Tracer recovery is rather random. Therefore firn water measurements were done only in the upper part of the glacier, were discharge amount are assumed to be lesser. There is no possibility to measure how much tracer is passing without being detected. Therefore only time-path measurements were carried out.

Due to high radiation input and short distances, Fluorescein is not suitable. Fluorescein is decomposed by light rather fast. Sulphorhodamin G and Naphtionate are of lesser light sensibility. The short measurement distances favored the latter substances.
Previously to the experiment the slope angle was calculated by a perpendicular and a measuring staff. Then the slope line was estimated for installing the fluorometer. In a first step a rectangular hole was digged onto the firn. Into the upper firn layer was hacked a gully of approximately 1 m. The gully's purpose was to collect the relatively small amount of water flowing in the basal saturated snow layer. The field fluorometer can only work if enough water is conducted through the optical cell. Intruded air leads to a measurement blackout. In addition, drainage was estimated by filling beakers and measuring the elapsed time.
The pre-soluted dye tracers were sucked up by the injection pipe and injected directly onto the firn with distances of 8, 15 and 25 m to the detection device. The presumption was that water flowing on the firn layer must be significantly faster than meltwater percolation through the snow layers.

Problems emerged because of deviances in slope line, especially with greater distances. Interesting site effects could be observed by measuring turbidity changes due to tracer injection and precipitation.

4.2.4 Supra-, en- and subglacial Tracer Tests

In order to measure travel times of water through the glacial body fluorescent dye measurements were done. The semi quantitative method of fluorescent dye tracing can specify time path relationships, as well as discharge measurements through tracer dilution.
The following experiments were carried out in two main measurement campaigns. The first was hold in the middle of the ablation season 2006, from the 26^{th} – 29^{th} of July, the second one at the very end of the ablation season, from the 3rd – 5^{th} of October.
For en- and subglacial pathways investigation Fluorescein and Naphtionate was used.

Fluorescein was used for longer distances as the substance has a low detection limit and is lesser disturbed by turbidity than other dye tracers. The initially used substance SulphoRhodamine G is not appropriate in subglacial channels. Water in subglacial channels transports sediments from glacial abrasion. The turbidity background signal on the outlet of the glacier can be very high (over 100 NTUs-Nephelometric Units).

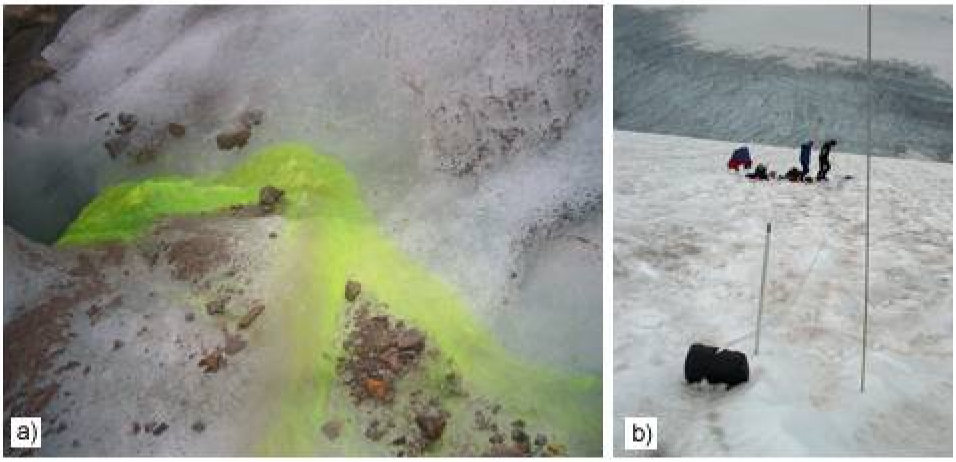


Fig. 15: a) Fluorescein release into glacier moulin at '*Oberer Boden*' ; b) Dye tracer release into basal snow layer at experimental site '*Lieslstang*'; injection pipe in the foreground, *Source: own picture*

A preinspection of the area disqualified not adequate injection points. Either they were not connected to the englacial system and not representative or simply difficult to reach or too dangerous.
Fluorescein and Naptionate was prepacked in the laboratory in portions of 20 and 50 g for better transportation since I was carrying out the experiment alone. The amounts were then dissolved in meltwater at the injection site in a measurement beaker. The dissolution had to be done cautiously not to contaminate the area before the proper injection.

The dye tracers were injected (with different distances to the detection device) into

a) moulins
b) crevasses
c) the subglacial stream or brook

The *Goldbergkees* has a rather complex drainage system. It is not always clear which part of the glacier contributes to a distinguished discharge. Moreover, the subglacial river quits the subglacial system before falling over a cascade of about 200 m and then to disappear under the glacier tongue. There is a smaller discharge system under the southerly glacier part, mostly including the glacier fall and the tongue. Though, it was hardly possible to carry out dye tracer experiments in this area, because of the huge crevasses area and the lacking of suitable injection points.

The fluorometer was installed on the outlet of the upper part of the glacier (*Oberer Boden*) and on the outlet at the glacier tongue, depending on the injection site (compare **Fig. 18**).

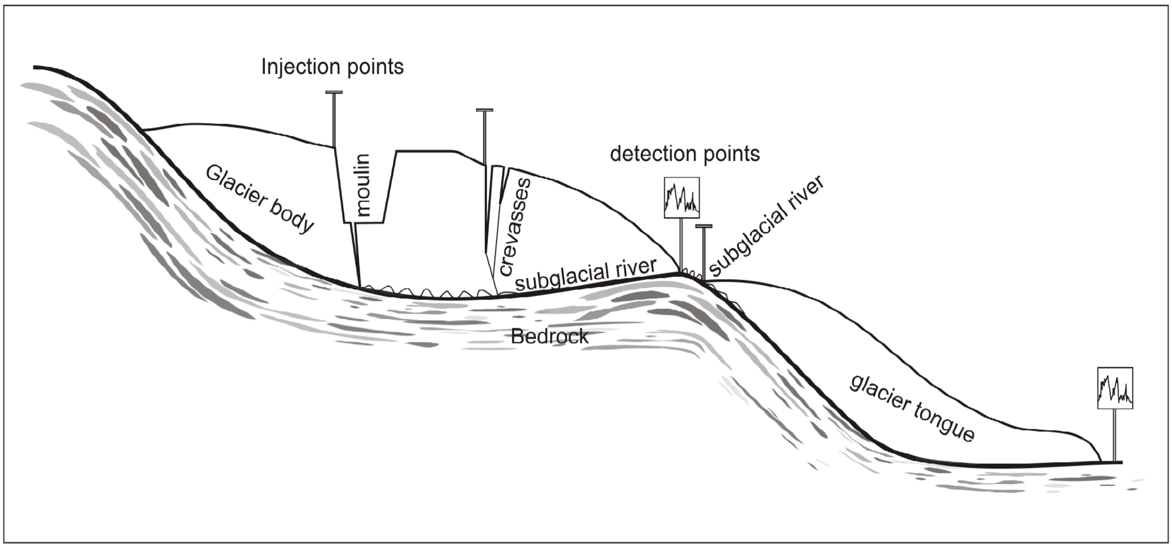


Fig. 16: Schematic figure, longitudinal cut of the investigation area with injection points and detection sites. *Source: Own illustration*

Detection intervals were set at 10 seconds. Registration simultaneously caught the Fluorescein- and the Naphtionate- signal. At the same time turbidity and temperature was measured.

The read-out of the data from the detection device occurred daily, immediately after the experiment end and not after the complete measurement campaign. The advantage is a immediate control of the measured data and the possible adjustment for further experiments.

Problems emerged for the unknown factor of dilution of dyes and flow path distance. Furthermore turbidity superimposed the dye tracer signal. In the lower signal strength of the measured dye, turbidity can cause exceeded noise which makes a calculation of the actual tracer concentration difficult

4.3 Adaption of different Dye Tracer for Field Studies on the Goldbergkees

In this field study have been employed three different types of fluorescent dyes. They have been chosen for special characteristics and were not randomized. The essential criterias were:

- Possible application for the snow body or for the water transport in ice
- Possible application in supra-, en- or subglacial channels
- Tracers must have different wavelength peaks for simultaneous measurements. Separation of tracer breakthrough curves can be done automatically by software
- Degree of light sensitivity in terms of decomposition
- Degree of acid sensitivity
- Low adsorbtivity regarding soluted sediments in melt water
- Price and availability
- Environmental and substantial harmlessness

The preliminary consideration of chemical and physical characteristics of various tracers ended up on three adequate fluorescent dyes:

a.) Fluorescein b.) Sulfo-Rhodamine G c.) Naphtionate

a) Fluorescein (Sodium-Fluorescein), (3-3-Hydroxy-6-oxo-9-(2´-carboxylphenyl)-xanthene): The most common used fluorescent dye in tracer hydrology. Vantages are its low detection limit (in clear waters under 0,0001 ppb) and its poor adsorbtivity. Fluorescein is light sensitive and does get decomposed rather fast. Wernli, 2003; gives application times for field studies in sunlight not over ½ hour. In this study, travel times were over half an hour, however it was used to investigate in en-and subglacial channels not being exposed to radiation. On supraglacial channels measurements with Fluorescein are hardly possible. Decomposition of light sensitive dyes is even stronger on glaciers as on not glaciated areas. The increased radiation has to be traced back on the normally high absolute elevation for temperate glaciers as well as the albedo and backscattering characteristics of bright surfaces. Fluorescein reacts sensitive to pH-value shifts. The fluorescence is best between pH-8 and higher. With pH-values around 6.5 the fluorescence of Fluorescein is diminished by 50%. The acidity is reversible but accelerates adsorption under pH-values of 5.5. This causes irreversible adsorption of cations by negative charged sediment compounds. *(Wernli, 2003)*

b) For this field study Rhodamines were used even if their application is problematic and infrequent. Rhodamine is easily adsorbed by organic matter and sediments but hardly decomposed by solar radiation. Its insensitivity to sunlight opens new capabilities. For e.g. water running in supraglacial meltstreams is relatively clear (not during rainfall). Though other tracers would not be adequate for an application in direct daylight. For this reason Sulfo-Rhodamine G was used for experiments on the glacier surface and in the snow body. Sulfo-Rhodamine G is decomposed 50 times slower than Fluorescein and therefore applicable in sunlight. Moreover it is easily distinguishable from other dyes through the different spectrometric extinction/emission peaks. Notwithstanding its analysis can only be done half-quantitative because of the poor adsorbtivity characteristics.

c) Thirdly Naphtionate was used for similar purposes as Rhodamine. Naphtionate (Naphthioneacid Sodiumsalt, 1-Naphthylamine-4-sulhoracid Na-salt) is a common dye tracer in hydrology but shows some negative characteristics. Naphtionate emits in the blue band. Depending on the clearness of water it simulates a distracting background-signal often difficult to interpret. This characteristic leads to higher detection limits. The detection limit of naphtionate is about 0.2 ppb. This means for the same emission rate we need at least 200 times more. Moreover its extinction/emission rate is generally only 18% of Fluorescein and leads to further problem in terms of suspended sediments resulting in higher turbidity. The detection can be interfere by a serious of parameters. Naphtionate is used thus only for short distances when turbidity is relatively low. Wernli, 2003; suggests distances of about 200-300 m.

4.4 Hydrological Tracer Test with the Flow-through Field Fluorometer GGNU- FL

Hydrometric water samplers are common as a portable or permanent measurement equipment. Nevertheless they show some disadvantages. The Geomagnetism Group of the University of Neuchatel (GGUN) developed a new down borehole fluorometer for field purposes. The flow-through fluorometer can measure dye concentrations immediately without the need to collect the samples mechanically and reanalyze them in the laboratory. Latter causes often contamination of the sample *(Schnegg and Kennedy, 1998; p. 48).*

The fluorometer possesses an optical cell with three superbright LED's. The light wells have a wavelength of 370, 470, 525 nm. This allows the device to clearly distinguish three different types of dye at the same time without disturbing the measurement among each other. The visible band spectrum is therefore fully covered. Moreover it is possible to measure turbidity at the same sampling interval. Turbidity is measured in nephelometric units (NTU) in a range of 0.02 to 400. The sampling intervals are of 6, 10, 30 seconds and 1,2,4,5,10,15 minutes depending on the number of simultaneously measured dyes *(Schnegg and Kennedy, 1998; p. 48).*

Fig. 17: Flow-through field fluorometer GGNU-Fl and detection sonde. *Source: own picture*

The GGNU-FL fluorometer is electrically operated by simple low voltage lithium batteries operating 2-4 weeks. The fluorometer collects data and saves it directly to a memory card (Flash Card). For this reason the fluorometer can work automatically for days even weeks without surveillance. It can be analysed also immediately in field with a workstation laptop to visualize the results. Moreover it is light-weighted (only 17.5 kg with two batteries) and inexpensive. For glacier studies this may play an important role when it comes to transport and user friendliness. The appending software for application control is simple to use as also reliable. Nevertheless the fluorometer can be driven also manually without the need of a laptop. The control panels on the main control device operate the most important issues itself. The software is in this regard only for calibration, visualization and fine tuning as well as for automated recalculation of the tracer quantities.

The borehole sonde is a metallic cubical cylinder of approx. 5.5 centimeters of diameter. Therein is located the optical cell. The sonde is connected to the main device with an optional 10 meter cable. It allows the insertion in boreholes in the glacier ice. The advantages of a cable connected sensor are obvious. It was possible to insert the sonde into crevasses or badly accessible sites.

Some problems can be caused due to water bubbles in the sonde. Especially in turbulent waters, bubbles can invade the sonde and interfere the measurement process. To avoid the entry of bubbles I provided the sonde water entry with a funnel and filter. The filter has been tested previously to guarantee operation in field afterwards.

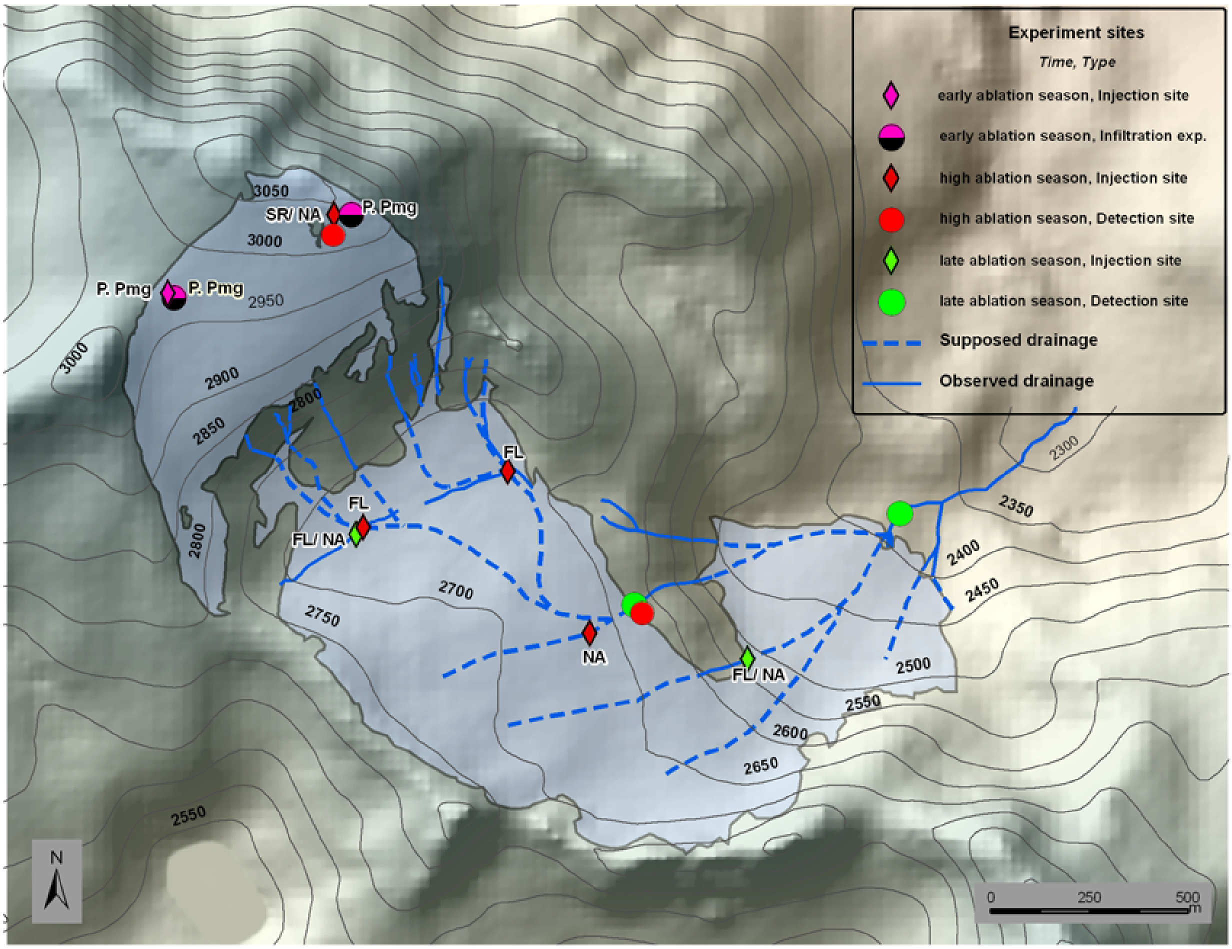


Fig. 18: Glacier area and field experiment positions. The supposed drainage pathways shown as dashed line are visually deduced from GIS analyses of glacier thickness measurements with GPR (Ground Penetrating Radar) and digital elevation models after Binder, Brückl, Roch, Behm, Schöner, 2009 in press. (For further details go to chapter 6.1) Glacier extension digitized after aerial image 2003.

5 Data Base

5.1 Primary Data

5.1.1 Photographic Documentation

Local conditions and circumstances where documented photographically in the run-up of the field experiments. The purpose was to select the right experimental sites as well as to collect basic information about flow processes. This step was important regarding the feasibility of the future field experiments.

In all further experiments (snow, firn pathways and glacier channels experiments) pictures where taken to register the intermediate steps of any experiment.

5.1.2 Coloured Dye Application and Injection on snow / firn

The extracted data of the coloured dye experiments was rather qualitative than numeric or quantitative. On the basis of observation and time to path relationships, precious information could be extracted. The collected data of this experimental step is to be regarded rather supportively. It can explain obscurities of results in later experiments.

5.1.3 Fluorometric Data

With the assistance of the down borehole field-fluorometer GGNU quantitative data about fluorometric dye tracer concentrations could be collected. The fluorometer could measure in pre-selected intervals of 6, 10, 30…seconds.

Dye concentration is measured in Micro-Sievert (mS) and calculated subsequently by the included software the measured concentration, typically in parts per billion (ppb). If calibrated properly, calculated concentrations (ppb) are precise.

Simultaneously turbidity, indicated in nephelometric units (NTU); and temperature indicated in degree Celsius is measured.

Tab. 2: Data structure of fluorometric dye test measurement with the GGUN-FL-Fluorometer

GGUN-FL	Concentration	Turbidity	Temperature
Time	microS/cm	NTU	T°C
06/07/27-14:46:31	0.01	110.58	0.97
06/07/27-14:46:37	1.03	109.89	0.97
06/07/27-14:46:43	3.67	105.56	0.97
06/07/27-14:46:49	4.37	100.76	0.97
06/07/27-14:46:55	4.26	109.27	0.97
06/07/27-14:47:01	3.71	96.18	0.98
06/07/27-14:47:07	0.01	97.17	0.98
06/07/27-14:47:13	0.01	110.74	0.98
06/07/27-14:47:19	0.01	104.55	0.98
…	…	…	…

The exact time and the date (resolution in seconds) are collected synchronically. The data is stored on a simple flash disc of 15 mb volume, accessible easily over a flash card reader. The data must be postprocessed with the included software to get the adequate concentrations. Nonetheless the original preprocessed file is accessible as a common .txt file.

Due to relative short experimental times resulting in smaller data quantities it was convenient to adjust to the highest time resolution (6 seconds-intervals for a single tracer breakthrough and 10 seconds- intervals for 2 or 3 tracers).

5.2 Secondary Data of hydro- meteorologic Monitoring

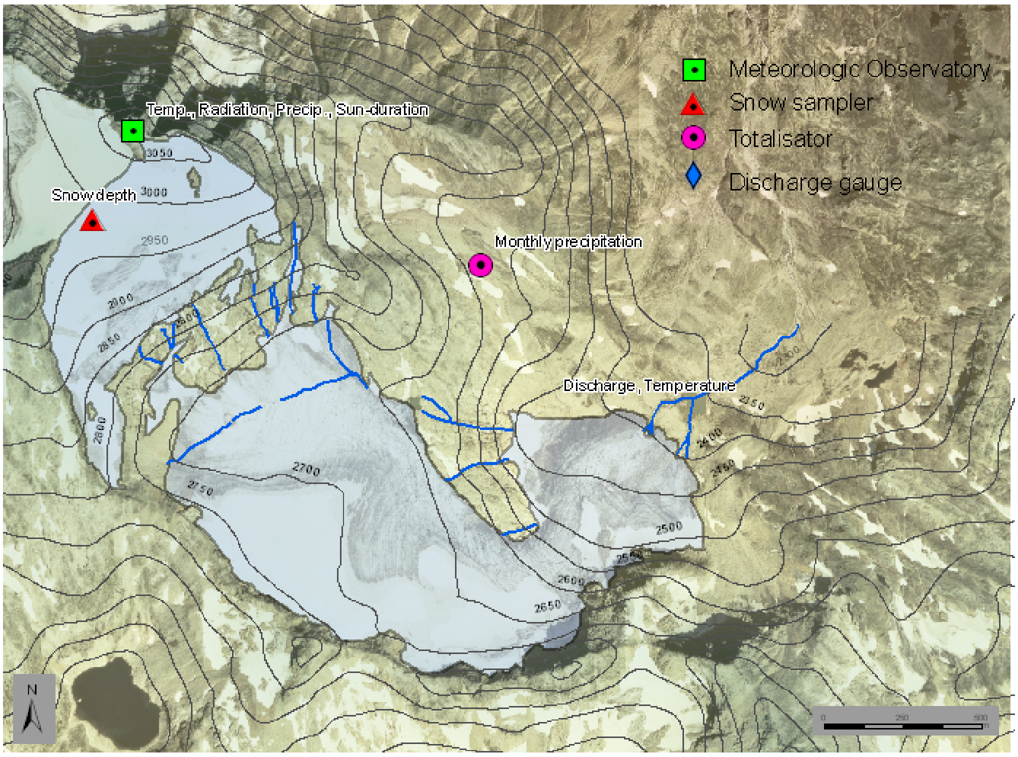


Fig. 19: *Goldbergkees* with relevant meteorologic and hydrographic stations. Glacier extension digitized after aerial image 2003

5.2.1 Precipitation and Temperature Measurement

Precipitation and temperature data are provided mainly by the near *Hoher Sonnblick* meteorologic *Observatory*. The permanently operated meteorologic station provides excellent meteorologic data.

There has been installed a hydrologic gauging station 250 m downstream of the glacier snout collecting precipitation and temperature data. Furthermore there is a totalisator collecting monthly precipitation data at 2580 m and within a distance of only 400m to the middle part of the glacier. The totalisator provides only monthly data but can be used to control the measured precipitation from daily or hourly registering stations. All hydro-meteorologic stations are shown in **Fig. 19**.

Tab. 3: Data structure extract of precipitation measurement for the Sonnblick Observatory data logger

Day	Month	Year	Hour	Preciptation in mm/h
...	...	...	...	...
8	8	2006	2	0
8	8	2006	3	0
8	8	2006	4	4,2
8	8	2006	5	9,79
8	8	2006	6	6,04
8	8	2006	7	0,74
8	8	2006	8	0,73
8	8	2006	9	1,54
8	8	2006	10	0,71
8	8	2006	11	0
8	8	2006	12	0
...	...	...	...	...

5.2.2 Radiation and Sun Duration Measurement

At the *Observatory* as well as at the hydrologic gauge station radiation is measured hourly. For this purpose (global) radiation is used. There can be significant differences between the *Hoher Sonnblick* and the ‘Oberer Boden’ or glacier tongue. The main areas of the glacier are affected by shadowing by the surrounding mountains whereas sun duration totals are greater at the *Observatory* because it is freestanding.

Fig. 20: Temperature sonde installed on the *Observatory Hoher Sonnblick* (3106m); Source: own picture

Radiation is measured in W/sqm. Radiation has the largest influence on the run-off process on glaciers. Temperature influences melt lesser than radiation. Radiation is linked to the sun duration, but can be still at a low level even if there is no sunshine registered. The clouds let still pass part of the radiation.

Therefore melt is significantly high on sunny and clear days, even if temperatures are on the lower part of the scale. The fact is though, that temperature reacts with a temporal retard to the radiation changes.

Moreover, the glacier microclimate plays an important role. Temperatures above the ice resp. in the first 20-30 centimetres are significantly lower in comparison with higher air layers or air above solid ground. There can be melt processes on the glacier ice, even if the temperature detection device is still registering temperatures under the melting point.

Tab. 4: Data structure extract of radiation measurement for the Sonnblick Observatory data logger

Day	Month	Year	Hour	Irradiance in W/m²
...	...	...	...	...
1	7	2005	7	0
1	7	2005	8	0
1	7	2005	9	14
1	7	2005	10	97
1	7	2005	11	175
1	7	2005	12	194
1	7	2005	13	231
1	7	2005	14	217
1	7	2005	15	203
1	7	2005	16	125
1	7	2005	17	14
1	7	2005	18	0
1	7	2005	19	0
...	...	...	...	...

5.2.3 Snow Height

On the upper part of the glacier a supersonic-snow-height sampler is installed. Latter measures the daily snow height, but misses in liquid precipitation. The sampler registers only the actual snow height without embracing the fallen snow itself. It must be computed by the subtraction of the measured precipitation on the *Observatory*. The snow sampler though is of great importance as it provides information about settlement processes of snow in winter and degradation of snow in spring and early summer.

The Central Institute of Meteorology and Geodynamics (Zamg) measures snow heights in continuous intervals for mass balance modelling. In addition snow profiles were made to calculate the snow -water value (density resp. amount of water per volume unit).

5.2.4 Water Gauge for Discharge Measurement at the Glacier Snout

Since 2005 a gauging station has been installed by the *University of Natural Resources and Applied Life Sciences* by *Dr. Gernot Koboltschnig.* The water gauge measures discharge near the outlet of the glacier through a pressure sonde installed at the outflow of a small lake at end of the glacier. The upswelling of the discharge amounts react in a rise of the water table and a gain in water pressure. The water gauge is operated throughout the whole ablation season, but can not collect data when snowfall and freezing in early autumn is beginning. Not until the beginning of summer the water gauge is set into operation.

Therefore a small, albeit important part of the drainage at the beginning melt season is not registered. The gauge station provides accurate data on a 10 minute scale and includes the whole glacier and two small side drainage areas.

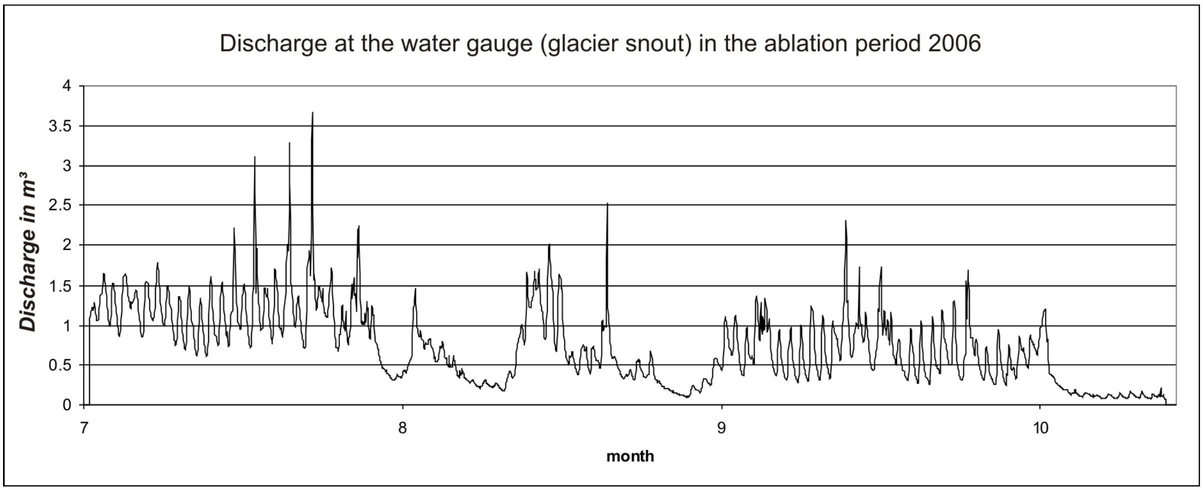


Fig. 21: Example: Discharge in cubic meters at the water gauge (near glacier snout) for the ablation season 2006

5.3 Secondary Data of Mass Balance Modelling and Snow height Measurements

Mass balance modelling data for comparison with annual meltwater amounts is provided of the *Central Institute of Meteorology and Geodynamics (Zamg).* By using the installed ablation stakes, snow depth measurements and from geographical interpolation over the glacierized area, annual melt amounts (snow –water equivalents) are calculated *(Koboltschnig, 2007).* It is possible to draw conclusions from the annual water balance to the annual mass balance of glaciers (see subchapter 2.1.5).

Nevertheless it is of lesser importance for the analysis of the consecutive results.

5.4 Secondary Data of Remote Sensing and GIS

Data from remote sensing was used for Gis applications and general geographical purposes. For geographical purposes in GIS a precise and high resoluted Digital Elevation Model (DEM) is needed. It is the basis for further calculations. For the positions of experimental sites as well as positions of various data samplers in the region maps or better orthophotos are needed.

5.4.1 Orthophotos

In 1998 and in 2003 aerial surveys were carried out over the Hohe Tauern region, collecting data about the glacierized areas. The orthophotos are available in a resolution of 0.5 meters (over ground). The outline of the glacier shape though was generated from the aerial image of 2003 and therefore includes part of the great length losses of that exceptional year. For drawings and maps though the older version of 1998 was taken since the actual version was taken after snowfall. However both images deliver useful geographical information about actual size, the outline of the glacier body, eventually also indices for mass reduction, the position of crevasses, snow and firn distribution and discharge pathways *(Koboltschnig, 2007; pp. 50-51).*

5.4.2 DEM and Computed Data

The "Digital Elevation Model"- DEM, is available in a resolution of 10 meters. It covers the whole Goldberggroup. The DEM has been generated out of stereoscopic images.

6 Results and Discussion

6.1 Observed and Computed Drainage

Discharge on the *Goldbergkees* is generally highly fractured. Not only is it split into different basins, but basin 2 must be subdivided into two topographically differing subareas as well.

Between *Oberster Boden* and *Oberer Boden* glacier movement generates the glacial deepening by abrasion leading to steep rockwalls. We can expect glacier ice to be thinnest were it flows fastest. Ablation over the last decades led to a break-off from the upper part. The uppermost part then does not move fast enough to close the emerging gap. Discharge from the upper part (*Oberster Boden*) is highly segmented. It is not possible to detect a unique flow path. There are at least 4 different discharge pathways observable. None of these is flowing visible on the surface, but are covered by avalanche or wind deposited snow. Even if topographical characteristics point to a main drainage in these rinulets it is not possible to estimate flowing water amounts. Moreover there are other observable discharge pathways on the lowest not glacierized part of the *Oberster Boden*. A vast number of small channels is flowing in rock cracks and directly on the abraded rocks.

Fig. 22 shows observed channels on the *Goldbergkees*. The channelling system is highly fragmented were they are observable and superficial. On *Oberer Boden* supraglacial channels are observable. They mostly follow topographic characteristics. We can expect though englacial and especially subglacial channels to follow the topographic characteristics of the bedrock once they reached the glacier bottom. Therefore observable and supposed flow channels differ highly. Even if we do not know the exact positions of en- and subglacial flow channels we can deduce some suppositions. Binder et al., 2009 carried out measurements with ground penetrating radar (GPR) to register ice thicknesses. **Fig. 22** shows ice thicknesses for the *Goldbergkees*. By subtraction of ice thicknesses from the digital elevation model, an approximation of the bedrock topography is possible. It allows us to roughly estimate positions of subglacial flow paths. This approximation is not verifiable but can help understand en- and subglacial tracer experiment results.

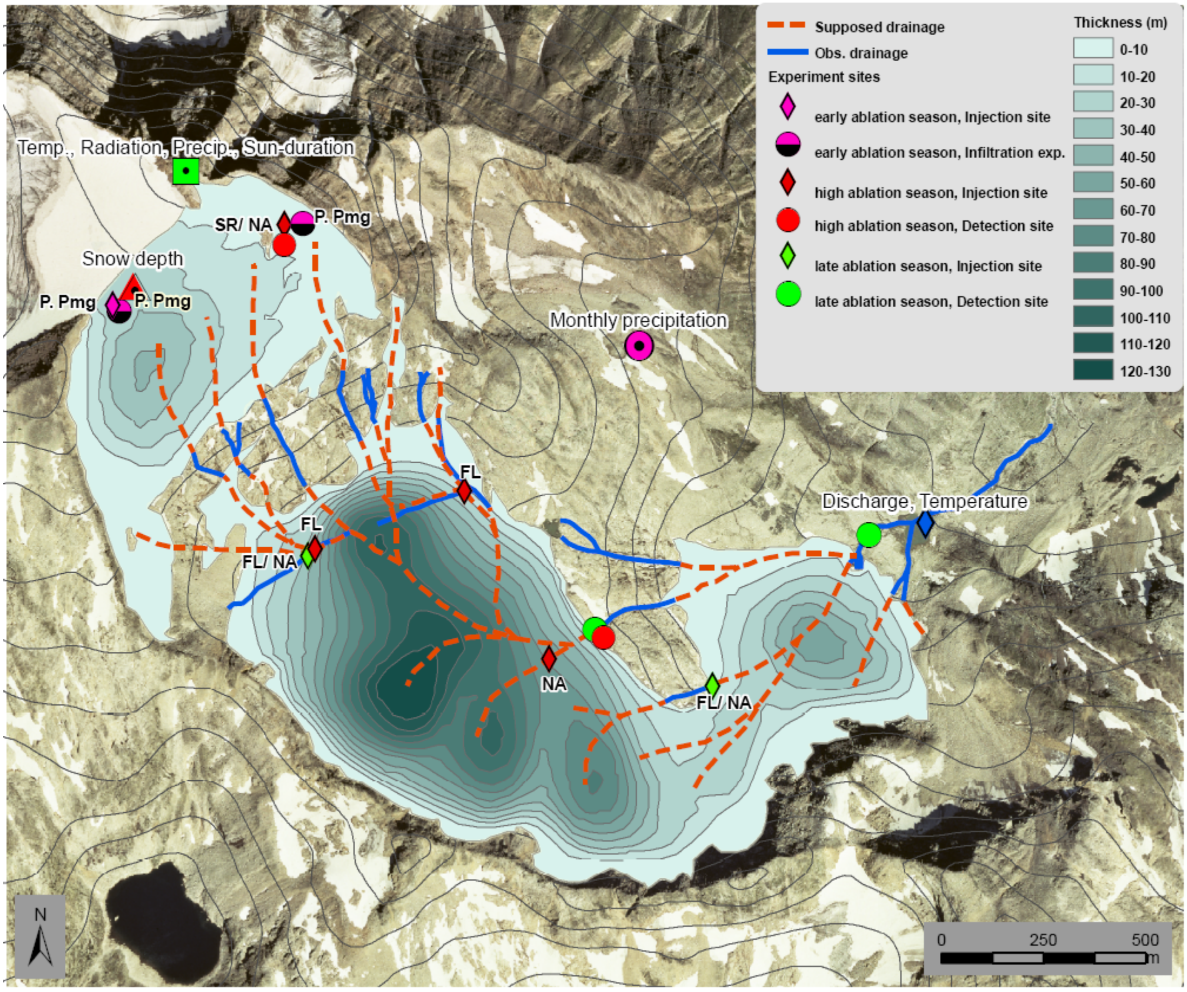


Fig. 22: *Goldbergkees*; Observed and supposed drainage channels after ice thickness overlay; experimental sites and ice thickness. Ice thickness after Binder et al. 2009 in press.

We can expect the englacial conduits to reach sooner or later the bedrock. But not all of the englacial passages will reach the bedrock. They can continue horizontally and never touch ground. Sediment analysis could substantiate these suppositions but were not done in the course of this investigation. According to the ice thickness measurements there is a bowl like form of the topography in the deepest recesses. Water could be collected subglacial leading to englacial or subglacial lakes or water holes. There is no evidence for a subglacial lake but the poor tracer recoveries for Fluorescein measurements in this area point to such an existence. The results for the late Ablation test campaign could be better explained with the existence of water fills but may have other reasons as well.

Moreover the spatial resolution of the ice thickness data is too low to calculate exact topographical models without the ice cover.

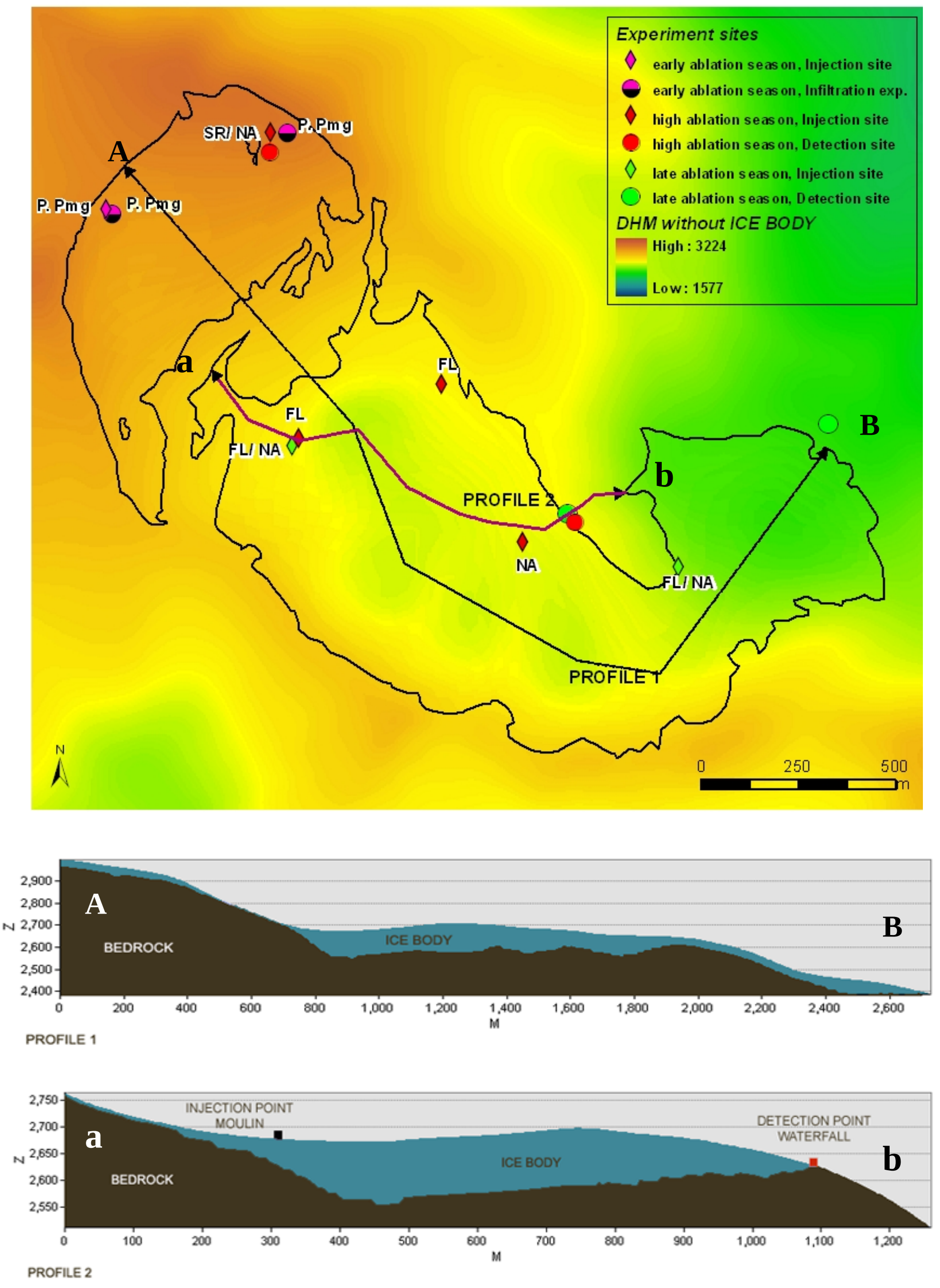


Fig. 23: Computed topographic model of the *Goldbergkees* without ice overlay (DHM – ICE THICKNESS= BEDROCK): Profile 1 cuts the glacierized area beginning at the *Observatory* over the *Oberer Boden*, following the glacier fall to the tongue; Profile 2 cuts the glacierized area following assumed flow paths for an injection into the glacier moulin ending with the bottom of the glacier fall. The Profile graphs at the bottom are computed by subtracting ice thickness data from the original DHM (real topography). Ice thickness data is derived from Binder et al., 2006

Due to the fact that subglacial channels follow bedrock topography if they reach it flow length deviances in relation to shortest distance flow are obvious. Measured ice thickness and the deduced flow paths for subglacial channels underline the division of the glacierized drainage basin.

It was possible to observe different discharge quantities, but we do not know amounts and especially velocities of water flowing in the main discharge channels. It was not possible to locate an appropriate test site for englacial drainage in this area. An estimation of meltwater flow paths was not possible. Moreover discharge amounts for accessible channels were too small for a continuous field fluorometer test. Even if tests should have been possible it would have required a high amount of samples to investigate local flow conditions.

6.2 Snowpack Investigations with the Coloured Dye Potassium Permanganate

6.2.1 Aerial Dispersion of Coloured Dye on Snow

Precisely at the beginning of the ablation season, coloured dye was dispersed on the snowpack to investigate percolation effects of water in the semi-frozen snow overlay. The visual analysis is not really precise in a mathematical quantitative context but redraws the pathways, water follows percolating through an inhomogeneous snow overlay.

6.2.1.1 Experimental Site 1 (Ultrasonic Snow Depth Sampler)

The first experiment was carried out on the 12th of June at 19:00 spraying 100 ml (6,5g dried matter) of a saturated Potassium Permanganate solution on an area of one sqm at the experimental site near the ultrasonic-snow depth sampler (**Fig. 18**).

The excavation done after approx. 17 h following the slope on the western side of the marked area resulted in a total length of 4,5 m till Potassium Permanganate tracks were not identifiable any more.

After another 20 hours on 14th of June at 8:00 a.m. the left, eastern side of the profile was digged out to measure further developments. The potassium permanganate plume advanced another 1.5 m until it could not be detected visually any more.

In both profiles horizontal displacement of the dye was greater than the vertical percolation. Nevertheless in some parts of the profile vertical displacement was clearly observable.

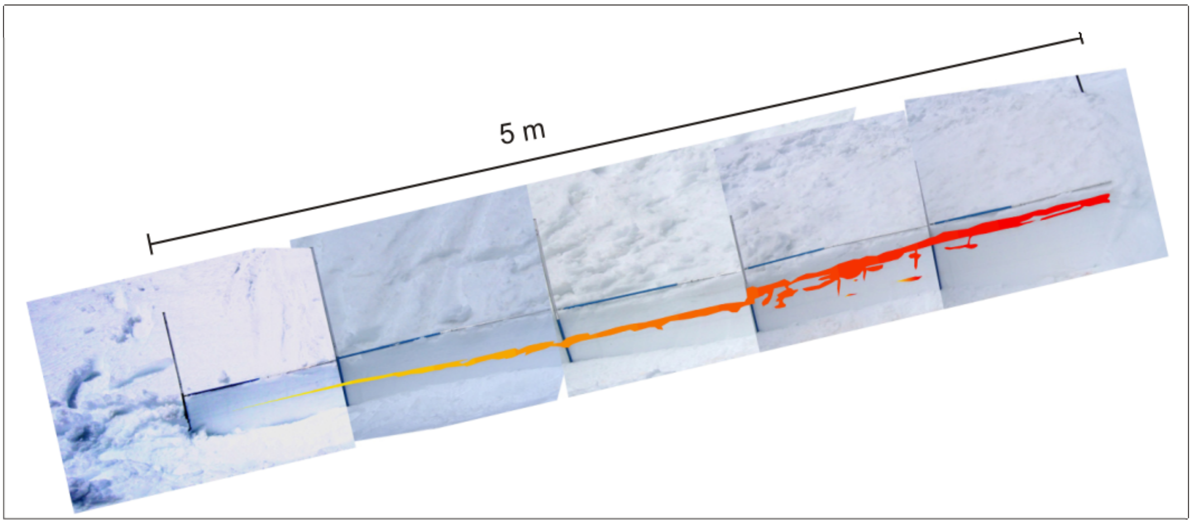


Fig. 24: Cross section on experimental site 1 (Ultrasonic Snow-depth sampler); first excavation after 17 hours; dispersion area (first section on the right side). Concentration decline is visualized by more yellowish colours. Graphics angle reflects natural inclination. Source: own picture, reworked.

The second excavation on experimental site 1 showed similar development. The dye plume proceeded for another 1.5 m further. There were no local vertical intrusions observable although digged only one meter westward. This is an indication for the spatially limited development of vertical flow channels. Generally the profile was much clearer and more uniform than the first. Nevertheless there are similarities. The diffusion plume flows mostly in the same depth. Moreover there is a similar drop-step approximately at 2 m for the first profile and 1,5 m for the second cross section. After approx. 6 m the dye plume was no longer visually detectable.

The progression resulted in 0,21 m h^{-1} for the first excavation and only 0,08 m h^{-1} for the second excavation if we assume that it progressed similarly in this part of the dye plume. It may be due to the fact that meltwater produced in the first section of the profile did not reach farther but may be also due to exceeded dilution (to small dye amounts). Nevertheless dye dispersion for experimental site 1 showed the relatively slow movement of water in pre-summer snow due to the high containment abilities of scarcely decomposed snow.

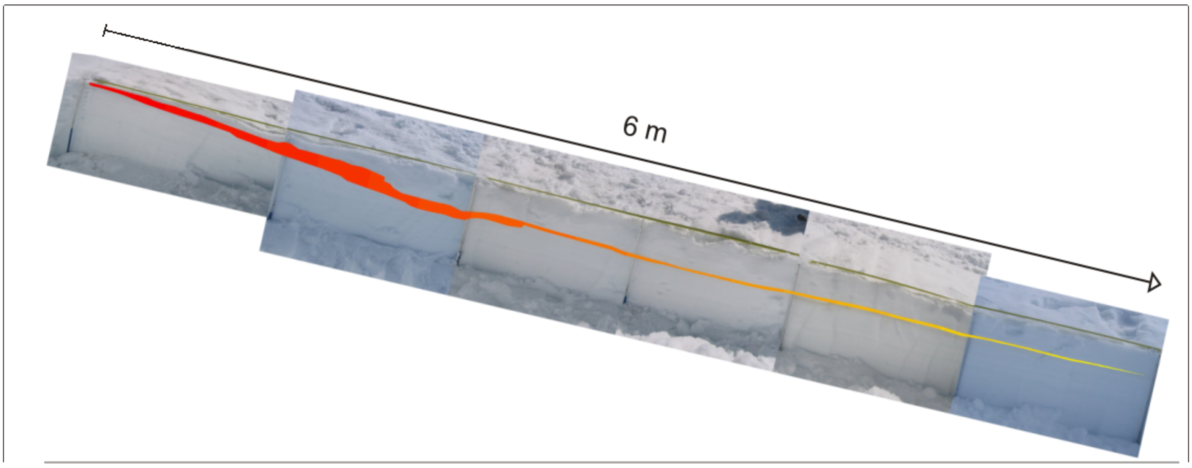


Fig. 25: Cross section on experimental site 1(Ultrasonic Snow-depth sampler); second excavation after 37 hours (17 + 20 h). Source: own, picture, reworked.

For both profiles similar densities in the different snow layers were measured. Moreover ice lamellae and high density layers could be located in the same depth.

According to **Fig. 26** there was a short melt period from the 26th to the 28th May. It could have caused an infiltration of meltwater into the first 10-30 centimeters. Especially in this depth higher density layers were recorded, due to later refreezing of infiltrated water. We can assume that the meltwater front could not intrude further into the snow layer. Particularly on the meltwater front snow is usually saturated. The colder layers lead to freezing by contact and seal the interspaces between the grains. Not until further heat transfer through intruding meltwater can displace this pond layer deeper into the snow. An anew drop of the temperatures stops the process and builds the mentioned higher density layers in the snow.

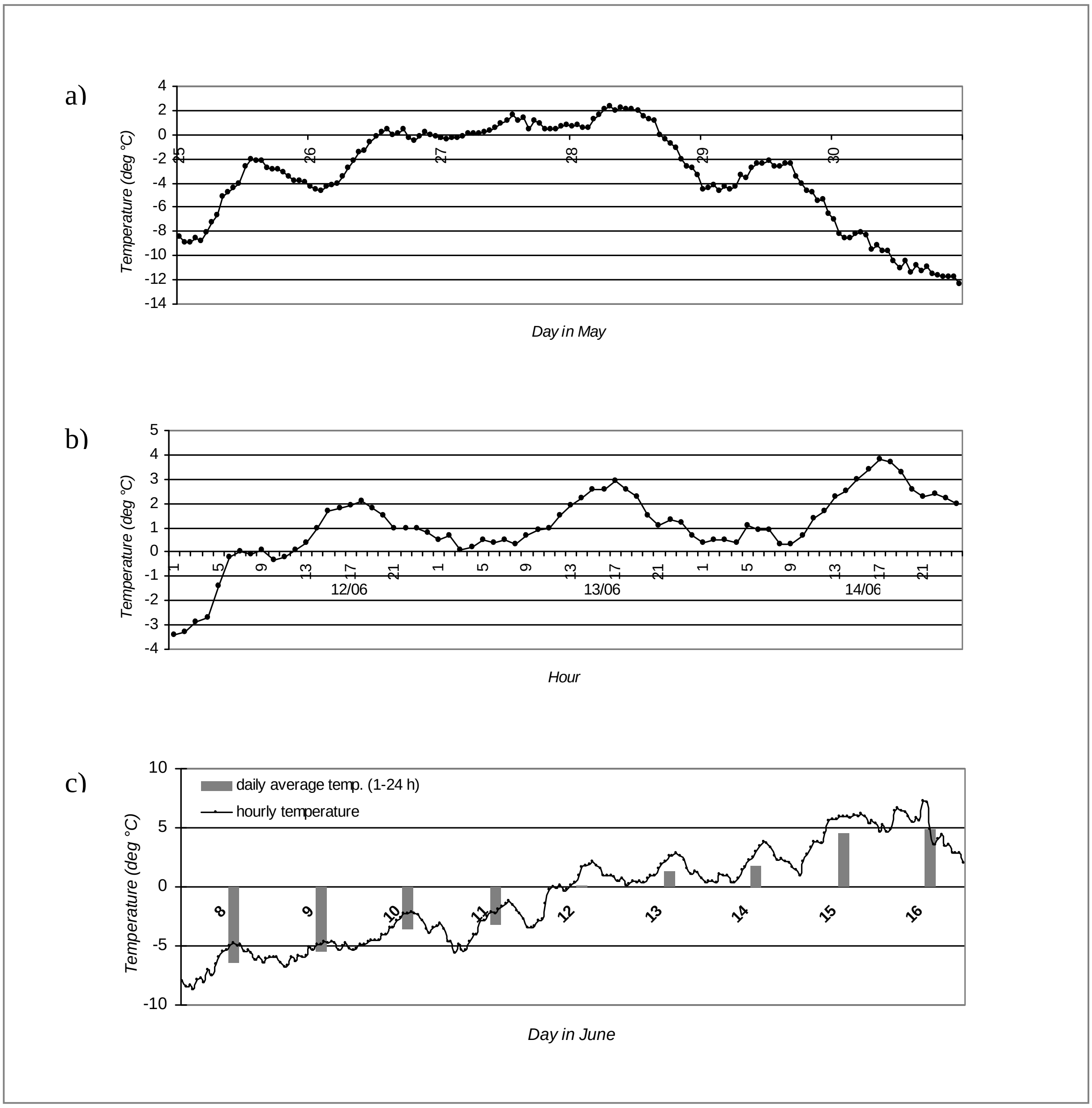


Fig. 26: Temperature record 2006 on *Hoher Sonnblick* (*Observatory*) at 3105m for a) the period of the 25th -30th May, the first short melt period in 2006, b) dye dispersion experiment period and c) the definite initiation of the ablation period.

6.2.1.2 Experimental Site 2 (*Lieslstang*)

At experimental site 2 namely *Lieslstang*, same quantities of dye tracer were dispersed on an equal area with similar inclination.

The dispersion took place on the 13th of June, one day after the initiation of the melt period. After 17 hours a cross section was created according to the criterias mentioned in chapter 4.2.2. A detection of the dye plume was possible for approx. 12 m, resp. 11 m of surface parallel flow path (complete profile- dispersion area). On different snow depths parallel flow was detected. **Fig. 27**, shows a cross section of the dye plume.

As in experimental site 1, there is a "stair" like intrusion into the snow body even if most of the dye flows on some denser snow layers.

It has to be noted that meltwater flowing horizontally on a broad front, intrudes vertically only in spatial restricted zones or channels. Therefore we can expect not to have recorded all vertical intrusions.

Although the cross section at experimental site 2 shows similarities in propagation depth of the dye plume there is a difference in the observed dispersion velocity. Whereas experimental site 1 displayed a movement of the plume by 3,5 m in the first 17 hours, experimental site 2 recorded a horizontal dispersion of at least 11 m. This is a rate of 0,65 mh^{-1}. Even the vertical intrusion was stronger in cross section 2.

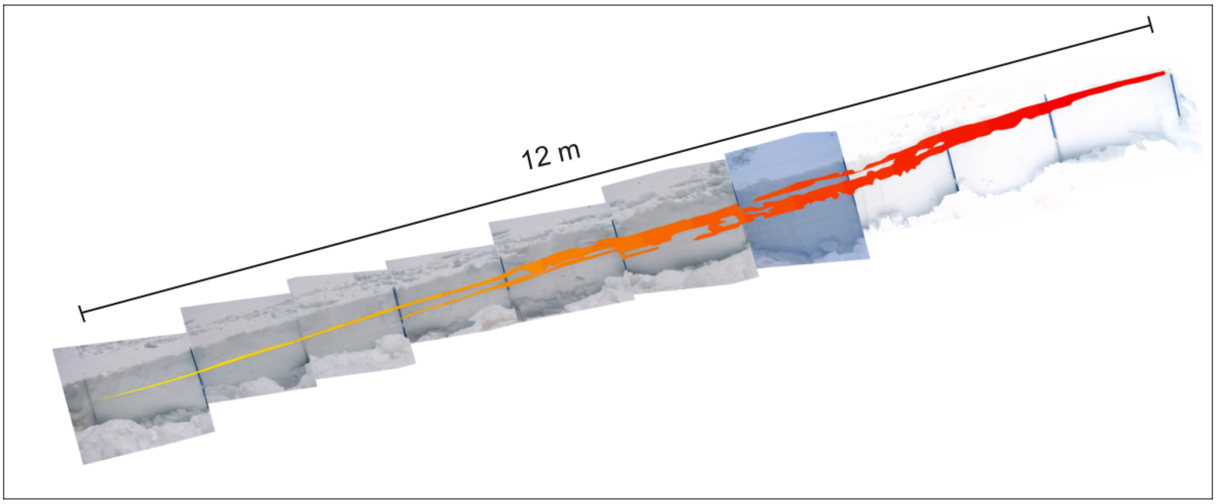


Fig. 27: Cross section on experimental site 2 (*Lieslstang*); excavation 17 hours after dispersion. Source: own picture, reworked.

Initially the dispersed dye follows the natural inclination and flows parallel to the surface on ice layers or at least denser snow layers. The denser snow layers are easily distinguishable by hardness tests and differ in colour. As long as the ice lamellae are connected water will flow relatively fast on the top of these hardly permeably layers.

Sudden breaks in the ice lamellae cause a vertical intrusion of meltwater. Depending on the crack in the ice layer still most of the water flows parallel to the surface. The break of the ice layers causes though a vertical displacement. After intruding deeper into the snow body water is forced to flow on a different structural level (see **Fig. 27**). In most of the cases it reaches another ice lamellae or denser part were the same processes recur. In other words water will channel through the snowpack reaching at last the locally permeable firn or ice layer.

The experiments cover only a special time frame namely the incipient ablation period. For the sampled snow layers, melt did not occur in this hydrological year before, if we let unconsidered a short melt period at the end of May (**Fig. 26**, a)). The snow layers did still mirror the wintery deposition. As shown in Tab. 5, temperatures were decreasing slowly by depth but varied only by 0,3 °C over the first two meters. The mentioned ice lamellae, even in deeper layers can be assigned to refrozen superficial meltwater induced by radiation melt. Nonetheless small amounts of meltwater can not reach the firn or ice layer as long as there are snow layers with temperatures under the freezing point. By intruding deeper into the snow it will refreeze on ‘cold’ snow grains. Not until the whole snowpack is equitemperate meltwater is mediated to the faster drainage systems. But even if water has reached the glacial system there can be found still insular parts in the snowpack with sub-freezing temperatures.

Tab. 5: Temperature sounding in the snow profile at experimental site 1

Snow Depth	*Temperature (deg. °C)*
0 -5	-0,1
5- 10	-0,2
10- 20	-0,2
20- 30	-0,2
30- 40	-0,2
40- 50	-0,2
50- 60	-0,2
60- 70	-0,2
70- 75	-0,2
75- 80	-0,3
80- 85	-0,3
85- 90	-0,4
90- 95	-0,4
95- 100	-0,4
100-200	-0,4

In cross section 2 at *Lieslstang* dye dispersion was faster and could intrude deeper into the snowpack. Temperatures overstepped the freezing point the 12th of June. Small melt amounts intruded the first centimeters of snow. In fact at the time of dispersion on experimental site 1 only the first centimeters were watery. The deeper and colder layers though avoided a further intrusion. *Campbell, 2006* speaks about 'textural horizons or layers of similar structure. Moreover we can expect water not to move until a certain limit of adhesion between the snow grains and the meltwater is under-run. Initially the meltwater in the snow can move only slowly but will accelerate as long as meltwater input does not deline. At cross section 2 meltwater movement was significantly faster according to this process. It could also infiltrate deeper into the snowpack shallow ice lamellae were breached by energy input through meltwater.

The vertical intrusion though is spatially limited. It depends on the maturity of the snow as well as local snow properties.

This process though is modified as the ice lamellae get decomposed. The growth of preferential flow paths or "flow fingers" in amount and diameter accelerates the mediating process as well as greater grain diameters. The decomposing metamorphosis of snow grains and the growth of greater one's at the expense of the smaller is highly accelerated by approximation to the melting point. Smaller snow grains retain water more effective than greater. With the progression of the ablation period grain structures get coarser. Horizontal movement of water in the snowpack decreases with the decomposition of the snow. At the end of the ablation period horizontal displacement of water plays an underpart in the mediation processes (**Fig. 28**).

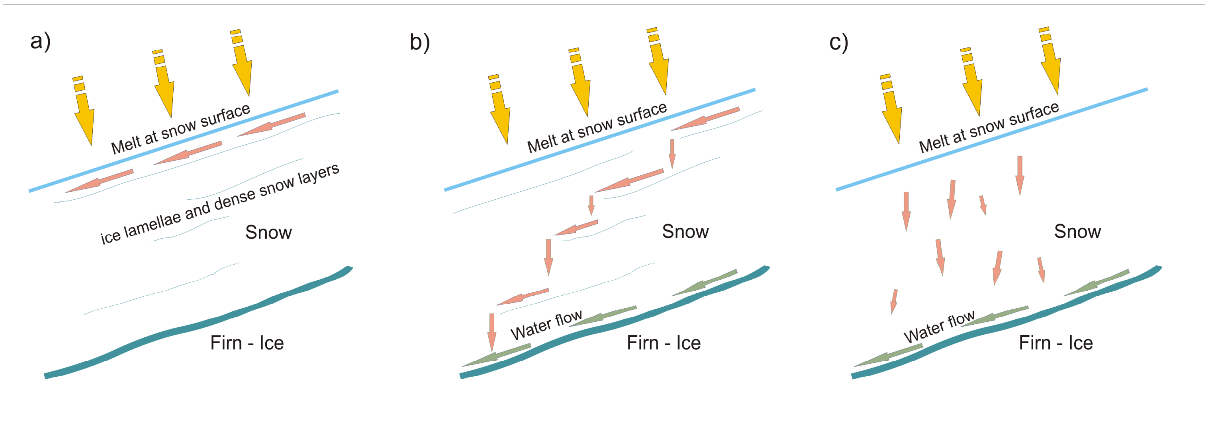


Fig. 28: Schematic figure, a) meltwater movement at the incipient melt season as observed in the field campaign from the 12th to the 14th of June, b) meltwater reaches firn or ice table (observed at lower elevations same time period), c) meltwater mediation in high ablation season (observed high ablation period field campaign). *Source: own illustration*

Assuming that snow is mediating water better in the high ablation season it is obvious that temporary retaining rates of meltwater are higher in the incipient ablation period. The present snow cover will deliver water slowly to the faster drainage systems.
In other words the drainage system in the snow cover of a glacier is getting more effective and accelerates. The sponge-like effect of a pre-summer snowpack is almost completely lost at the end the ablation season.

6.2.2 Direct Injection of Coloured Dye into Snowpack

To measure and delineate dye movement in the snowpack potassium permanganate amounts of 1,9 g diluted in water (saturated potassium permanganate solution) were injected through an injection pipe into the snow body.
According to previous investigations to snow depth and structure the injection was released in special zones of interest (over and under ice layers and in deeper areas). The deeper areas were still below the freezing point.
The injection was carry out the 12th of June, 20 m above the experimental site 1. At intervals of 50 centimeters 5 dye injections were carried out (**Fig. 13**).

The excavation on 14th June returned no dye tracer tracks. Although observable in the aerial dispersion experiments it was not possible to register any dye tracer in the excavated cross section.

A first supposition came to inadequate tracer amounts. To verify the assumption the experiment was repeated the next day but with thrice the tracer amount (approx. 5 g in 100 ml).
The excavation after 9 hours (exactly half the time of the previous experiment) returned only a very week tracer plume. Even if the solution was injected in a “sub-freezing point” (almost no free water) surround tracks were hardly visible.
Further observation showed that the injection process created a remaining shaft functioning like a local drainage. It channeled snow surface meltwater deeper into the snow and resulted in local above average drain. The melt waters of surrounding areas were channelized into the shafts. The cold temperatures in the deeper layers caused a sealing of the tubes walls. Therefore water was directed directly into the previous injection zone elutriating in short time the injected dye.

For this reason the experiment could not be implicated in this study.

6.3 Fluorometric Dye Experiments on Firn under Snow

6.3.1 Injection of Fluorometric Dyes on the Firn Surface

The following experiments were carried out on experimental site *Lieslstang*. The following reasons were relevant for choosing the test site:

- Intact snow cover
- No disturbance by crevasses
- Undisturbed surface (smooth firn surface)
- Easy delineation of slope line
- No excessive snow depth which complicates the injection and the installation of the fluorometer data logger
- Closeness to the meteorologic *Observatory*

The experimental site is only situated a few hundred meters away from the meteorologic *Observatory* providing excellent data for the sampled period. The slope has an inclination angle of 17-18 ° which is valid for most of the upper part of the glacier. The experiments could only be carried out on relatively short distances as exact flow directions of water under snow were not known and risk of missing the right sampling point was high. The purpose was to detect water flowing on the almost impermeable firn table. Even if small amounts of water can percolate through the firn body (especially in late ablation season as channels gets widened) most of the snow-induced meltwater flows laminar or channeled on the firn body resp. in the lowest snow layer. There is very little known about velocities with which water is moving under snow.
The dominant processes for water movement under snow at this time of the ablation season are shown in **Fig. 28**, c).

Surely the results can only give information about a very special period of the melt season and are strongly related to local geographic conditions and snow structure. Nevertheless it was possible to estimate characteristics and velocities of water movement under snow.
From firn excavations we know that water flowing on the firn surface is moving both laminar (plane) and channeled. De facto water is moving through the basal snow layer since the firn table behaves as a impermeable water table. The lowermost snow layer is characterized by a coarse porous snow structure. Dewatered, it shows a permeable structure with a tubelike small sized channel system. Samples revealed lesser adsorbtivity than snow from overlying layers. We can assume that the regular though stronger or weaker passage of water decomposes the snow structure.

The aquiferous layer on the firn body was easily distinguishable by the fact that there is almost no air included which is distinguishable in colour intensity. Whereas normal snow layers are bluish-white, aquiferous snow layers are of intense blue/gray colour. All cavities are air filled.

The amount of water flowing through this conductive layer is highly variable. It depends on the period of time in the annual cycle, daytime, melt intensity, involved melt area and whether water flows channelized or plan.
Estimations with a graduated beaker at 12:30 a.m. on 26th July taken in 10 second intervals collected over a section of 1 m width resulted in water amounts of approx. 0,052 l s^{-1} (+/- 180 l h^{-1}). The outflow in the excavated ditch was not perfectly uniform but showed an increase in the left third of the 1 m section.

6.3.1.1 Fluorescent Dye Tracer Tests on Day 1 (26th of July)

The dye tracer experiments were carried out the 26th and the 27th July. As the melt season started 42-43 days before around the 12th June, with a short temperature drop from the 30th June to the 2nd July, the snowpack was equitemperate for the whole depth.
In the experimental phase night temperatures were around 5,5 to 5 °C and around 10 °C during the day. Weather conditions were sunny with only partly clouded skies. Global radiation exceeded 800 W m^{-2} . For snow and glacial melt, radiation is the dominant. It contributes highly to snow melt. Temperatures play only a subordinate role. However rain on snow is of importance in the heat transfer from the surface into deeper snow layers. It can mediate energy faster and effectively into the snowpack whereas temperature increases influences only slowly deeper snow layers due to its low heat conductivity. In total high melt rates were observed in the experimental phase. In the evening of the 26th of July a thunderstorm caused a temperature reduction and heavy rainfall (**Fig. 29**)

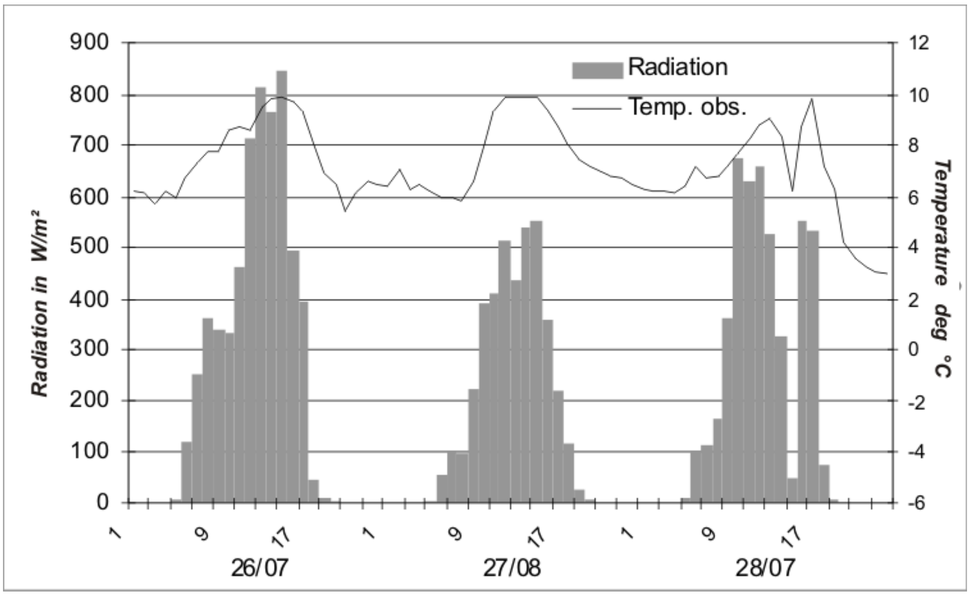


Fig. 29: Temperature and radiation on the *Observatory* for selected experimental period; temperature drop on 28th caused by heavy rainfall event.

A first dye injection was carried out the 26th of July at the experimental site *Lieslstang* at approx. 3020 m elevation. 1 g of Sufphorhodamine diluted in 30 ml of water was loaded into the injection pipe and released 3 m under the snow surface resp. directly into the aquiferous layer. The installed fluorometer at 8 m of distance recorded simultaneaously dye tracer concentrations in 6 seconds intervals. The resulting dye breakthrough curve is shown in

Fig. 30. It shows a steep rising limb and a pointed crest, followed by a rapidly falling recession (receding limb) with a slowly receding tail. Tracer dispersion is low as the distance to the injection point is only 8 m.

Results showed a first detection after 2 min 40 sec after the injection and a peak concentration at 5 min 10 sec. The signal is very clear and reflects maximum and mean flow velocities of water under snow. The calculated dominant effective flow velocity in this experiment is 0.0258 m s^{-1}, but mean flow velocities must be slightly lower. According to tracer hydraulics it is the point when 50 % of the tracer passed the data logger. After 5 min 30 seconds 50% of the tracer passed the data logger. Therefore the mean flow velocity or modal velocity was calculated 0,0242 m s^{-1} or more than 87,27 m h^{-1} **(Fig. 30).**

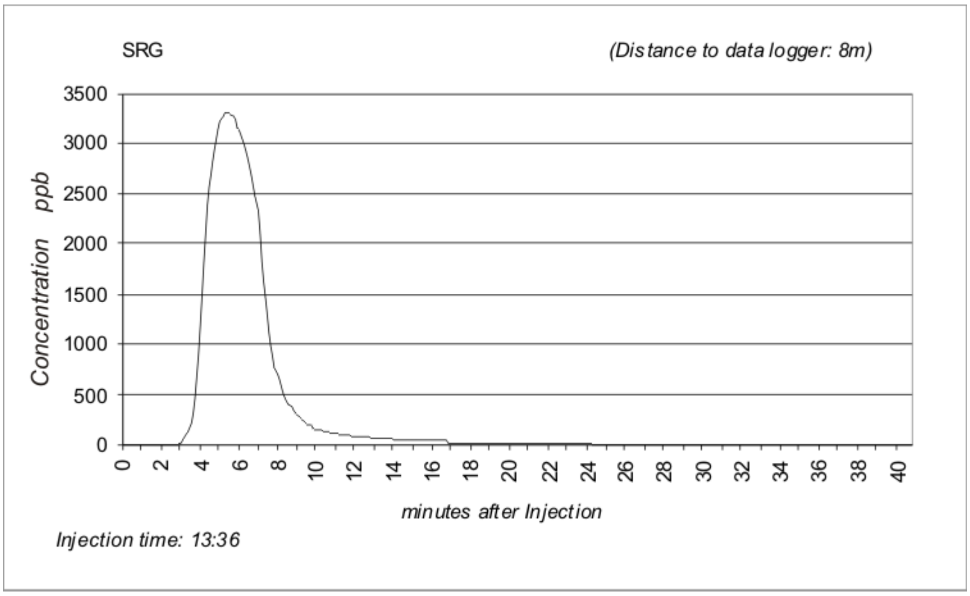


Fig. 30: Sulforhodamine concentration; injection on firn table the 26th July (experimental site *Lieslstang*); data logger distance from injection point: 8 m

Further dye tracer injections with Naphtionate on the 26th of July show similar results. Travel distance was 20 m. The breakthrough curve resulted clear but was slightly obscured by a background signal since naphtionate as well as Sulphorhodamine react sensible to turbidity changes. Turbidity rates for water flowing under snow are relatively low as long as snow reacts like a filter to turbidity-varying intrusions from the surface. But there is an empirical threshold value linked to the amount of infiltration meltwater. Higher melt rates can flush out solid matters from the surface and from inside the snowpack. To avoid an exceeding noise of the dye signal greater dye tracer amounts would be needed.

Fig. 31 shows a first dye detection around 6 – 6.5 min and a crest with a maximum dye concentration at around 8 min 30 sec. The recession is rapid but can not be analysed and compared with the previous experiment. The increasing background noise influences the dye signal. Although we can assume that 99 % of the tracer passed after approximately 20 minutes concentration values increased steadily for another 20 minutes, but with a 10-15% fluctuation of the mean value. Turbidity increased steadily in this phase. It may be related to high melt rates or a retarding passage of meltwater produced in the previous hours.
The calculated mean flow velocity over 20 m is 0.037037 m s^{-1} resp. 133.3 m h^{-1} and therefore significantly faster than in the previous experiment.
Melt is highest around 13:00 and 13:30 p.m. The retarding effect of the snowpack in mediating water to the firn table is not exactly known in this case but previous studies

yielded values from 0,17 -0.5 m h^{-1} for similar conditions (Campbell, 2006). If we sum it up to local snowdepths of approx. 3 m we get a retardation of at least 6 h. Nevertheless we can expect the values to be higher as structures in the snowpack were almost completely uniform. Snow structure decomposition has advanced to a grade were only coarse grains exist.

There is a jump in velocity from the percolation through ripe snow and the movement in the lowermost saturated snow layer. Due to higher pressures in the basal layer water is forced to flow faster. In theory there must be acceleration in the transit velocity when meltwater amounts increase. Simultaneously the vertical dimension of the saturated layer at the firn table increases. This will also get to higher flow velocities in the lower part of the snow covered glacierized area. The actual difference in flow velocity can only be estimated for this study.

Synchronically sulphorhodamine was injected into the basal snow layer (**Fig. 32**). To test needed tracer amounts, only a quantity of 0.1 g was injected. The registered concentration resulted not utilizable due to the influence of turbidity. Nonetheless there is a short amplitude possibly reflecting the breakthrough. This conclusion is supported by the equal incidence of elevated concentrations as in the previously described naphtionate test (see **Fig. 31**). For clear sulphorhodamine concentrations higher dye quantities have to be employed.

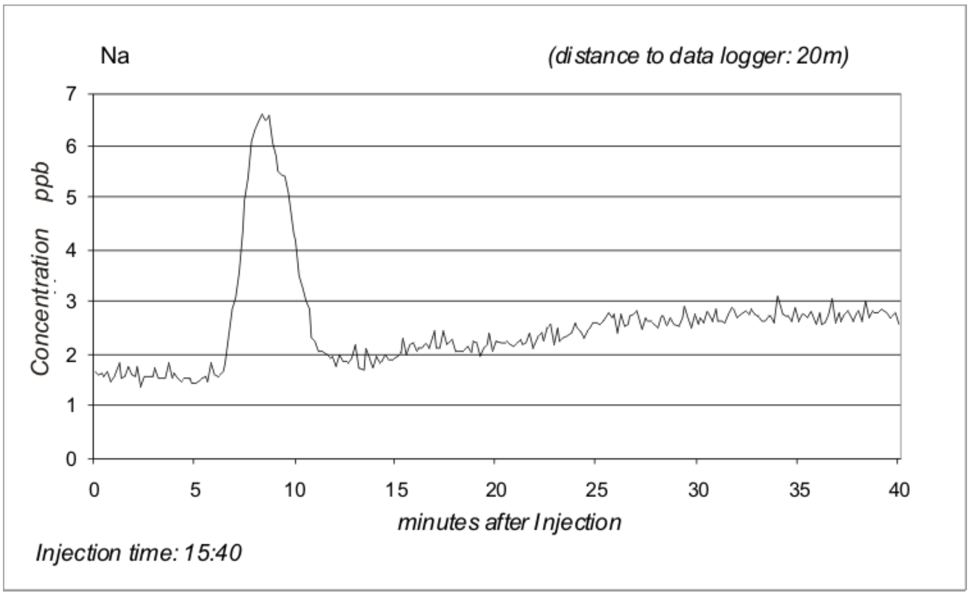


Fig. 31: Naphtionate conc.; tracer release under snow on firn/ice the 26th July (experimental site *Lieslstang*); data logger distance from injection point: 20 m

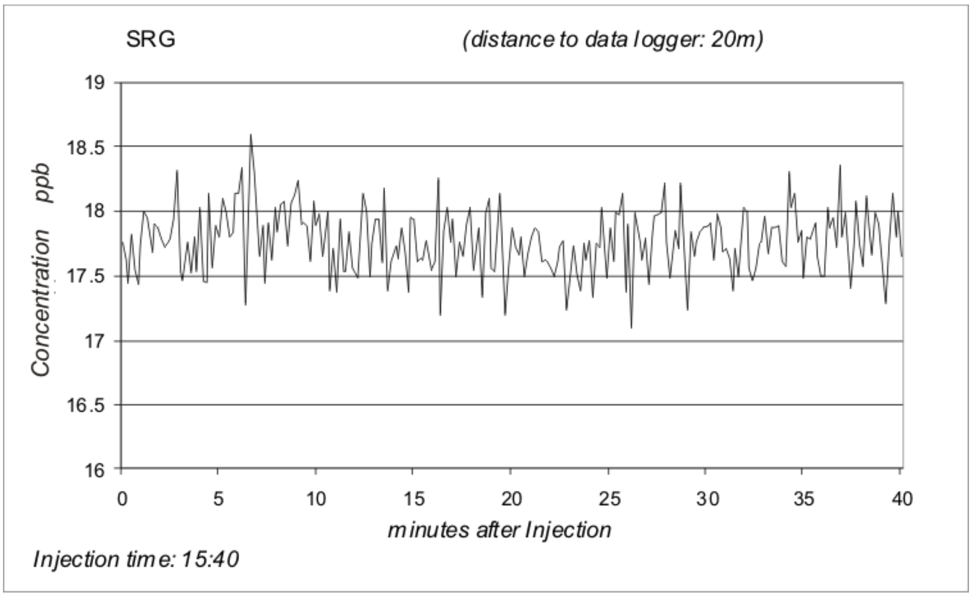


Fig. 32: Sulphorhodamine conc. for same experimental setting as Fig. **31** with very low dye amounts (0,1 g); background noise overlaps breakthrough signal

To collect data in greater time frames turbidity was measured over night. At 19:00 p.m. a thunderstorm brought heavy rainfall (see **Fig 33**). Precipitation samplers measured an hourly deposition of 14.49 mm from 19:00-20:00 p.m., followed by 2,158 and 0,219 mm in the next hours.

Turbidity raised simultaneously to an average value of 3.19 NTU and fell slightly to 2.63 NTU. From 22:00 p.m. to 1 a.m. it steeply increased to reach a sell at approx. 52.2 NTU. In the following 3 hours it fell back at 33.5 NTU then to rise again to a second crest at 5 a.m. in the morning. The slight increase in NTU with a peak at 20:00 p.m. can be assigned to the direct in-wash of particles and organic matter into the ditch excavated to collect water flowing on the firn table. The following and strong increase of the turbidity signal though can be assigned to the breakthrough of the rainwater. It infiltrated the snowpack and percolated through the snow layers then to flow downward on the firn table. Peak turbidity was reached at 1:00 a.m. in the night approximately 6 hours later than the deposition on snow. The second crest can not be explained properly, but could be caused by a second not registered local rain shower or from a flow path ramification.

The water movement through the overlying snowpack results in a retardation of 6-8 h, corresponding to approximately 0.4 – 0.5 m h^{-1}. It perfectly corresponds to values given by *Campbell, 2006*. Nevertheless the results explain only local conditions and can not be assigned to lower elevations with lesser snow cover. Moreover infiltration into snow at lower elevation can be faster as revealed in 6.3.1.1.

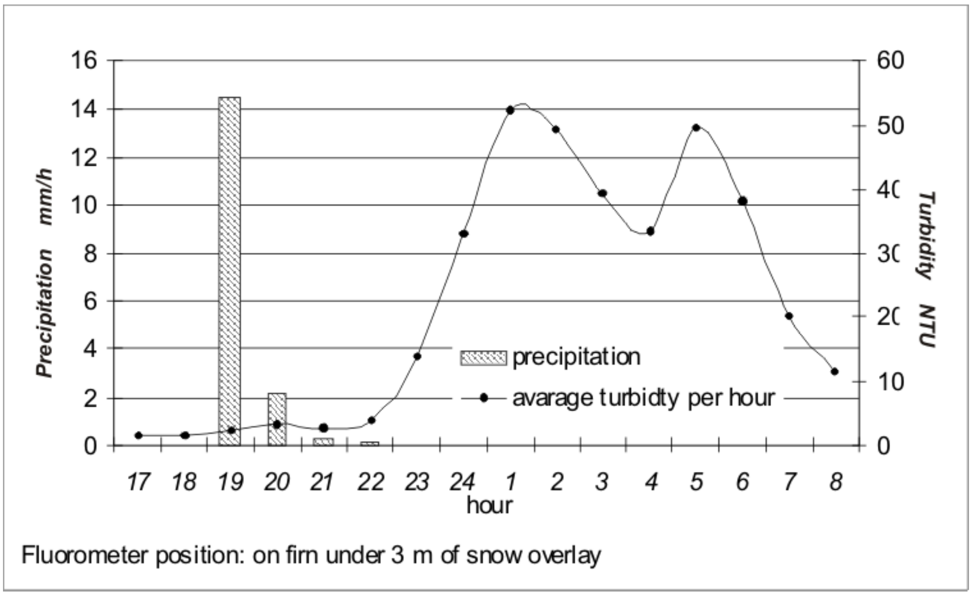


Fig. 33: Registered precipitation and measured turbidity at experimental site *Lieslstang* for vespertine rainfall event the 26th of July.

6.3.1.2 Fluorescent Dye Experiments on Day 2 (27th of July)

At 9:30 a.m. 1 g of naphtionate was released onto the firn table (**Fig. 34**). Records show first dye contact after 3 min 10 sec and a concentration peak after 5 min 50 sec. After 6 min 10 sec half the tracer amount has passed. Mean flow velocity is 0.0216 m s^{-1} or 77.8 m h^{-1}. The velocity is therefore significantly slower as resulted for same distances the day before. This could be because of lower flow rates in the morning affecting the flow velocity.

At 10:05 another tracer test was carried out. 1 g suphorhodamine was injected. Similar results emerged. First tracer contact was registered 3 min 30 sec after injection. Median breakthrough value was reached after 60 min 20 sec. Mean flow velocity was calculated 0.021 m s^{-1} or 75.6 m h^{-1}.

In **Fig. 35** the relation between turbidity and sulphorhodamine concentration is shown. Extremely high dye concentration affects turbidity and vice versa. This relation shows also the problematic application of sulphorhodamines in open sub- or preglacial channels. Sulphorhodamine can not be detected if turbidity exceeds certain values depending on the actual concentration.

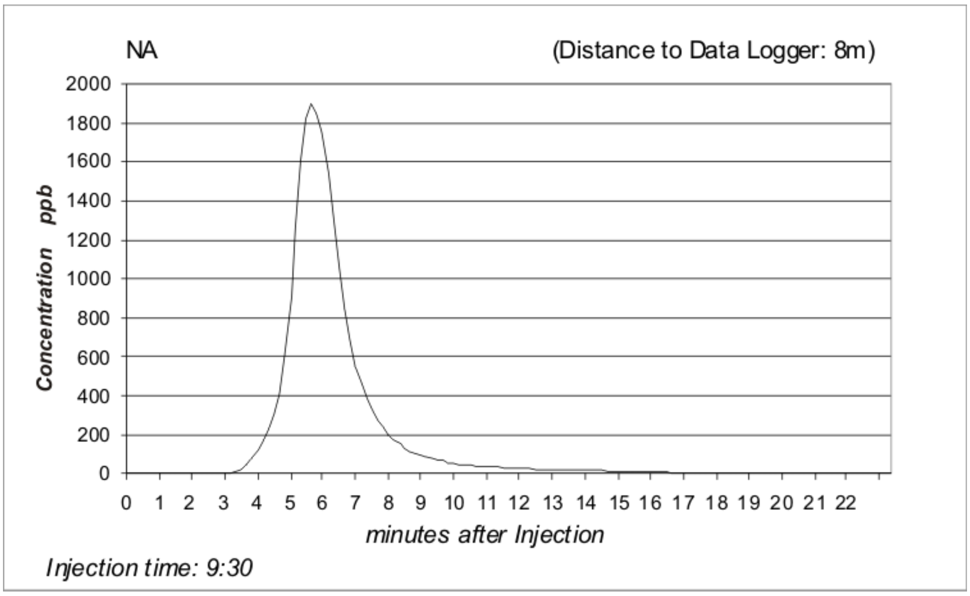


Fig. 34: Naphtionate conc.; tracer release under snow on firn/ice the 27th July (experimental site *Lieslstang*); data logger distance from injection point: 8 m

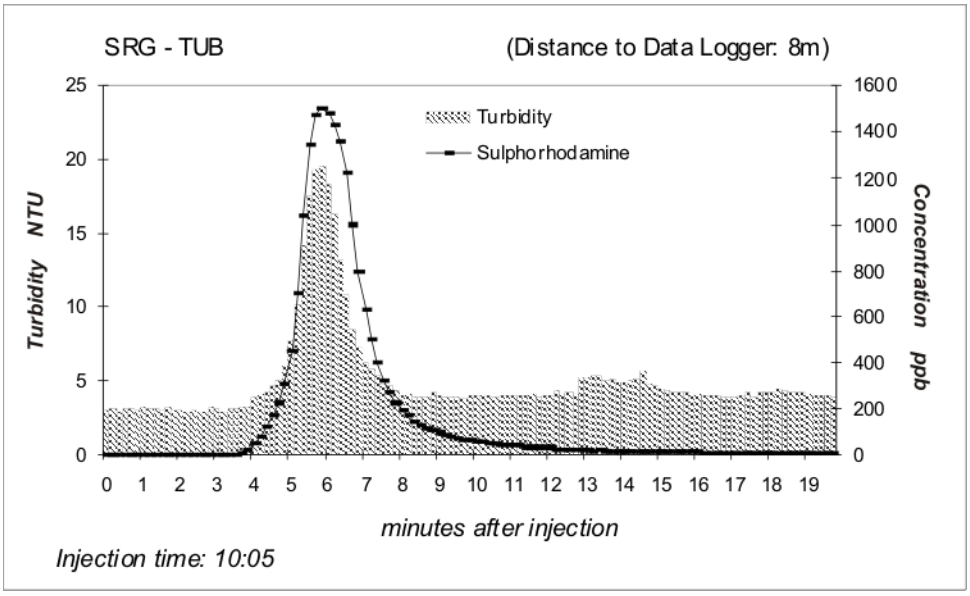


Fig. 35: Sulphorhodamine conc.; tracer release under snow on firn/ice the 27th July (experimental site *Lieslstang*); data logger distance from injection point: 8 m

A last firn tracer test was done at 10:40 the 27th of July, injecting 5 g of Naphtionate into the basal snow layer (see **Fig. 36**). First detection of naphtionate was registered after approx. 10

min. The rising limb of the breakthrough curve is not as steep as in the experiments before. It is an indication of the increased distance (20 m) or a higher dispersion. The crest is reached after 17 min. Mean flow velocity is calculated as 0.0196 m s^{-1} resp. 70.58 m h^{-1}. It is significantly slower than measured velocities for the 26th of July (133.3 m h^{-1}). Results from earlier experiments this day showed similar values.

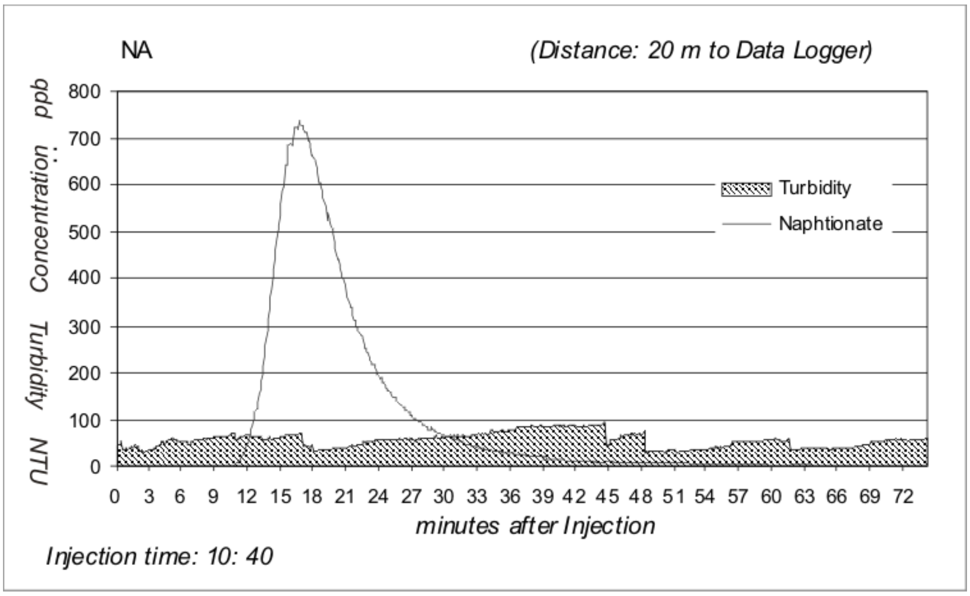


Fig. 36: : Naphtionate conc.; tracer release under snow on firn/ice the 27th July (experimental site *Lieslstang*); data logger distance from injection point: 20 m

All velocity measurements on experimental site *Lieslstang* show a wide range. Lowest measured flow velocities move around 70 m h^{-1} whereas fastest are around 130 m h^{-1} (see Tab. 2). There is a percentual increase in flow velocity of nearly 90% of the minimum value (see **Fig. 37**). We can expect not to have measured the real daily minimas or maximas. Therefore flow velocities can be slower or even faster. Conditions on the firn for this experiment were rather casual. Flow on firn could be more laminar or even more channeled. A laminar flow would decrease flow velocities whereas a channeled flow would increase flow velocities since the flow system is more effective in channels. Moreover it can be expected an increase in flow velocity from the top of the glacierized *Oberster Boden* to the bottom. It will not increase steadily but will reach an approximation value. Storage capacities for snow will decrease at the beginning of the ablation season faster than at the end but will reach a minimum at the end of summer. Retention is lowest when drainage through snow and movement of water on firn is fastest.

Nevertheless storage capacities (retention) of meltwater are better than in englacial or subglacial conduits. Although not exactly dimensionable for the whole area it was still possible to demonstrate the sponge like effect of snow cover by the past experiments.

Tab. 6: Mean flow velocities measured in field experiments the 26th and 27th of July. Travel distances varied between 8 and 20 meters.

Day	**25th July**		**27th July**		
Time	**13:36**	**15:40**	**9:30**	**10:05**	**10:40**
	87.27 m/h				
		~130 m/h			
			77.8 m/h		
				75.6 m/h	
					70.58 m/h

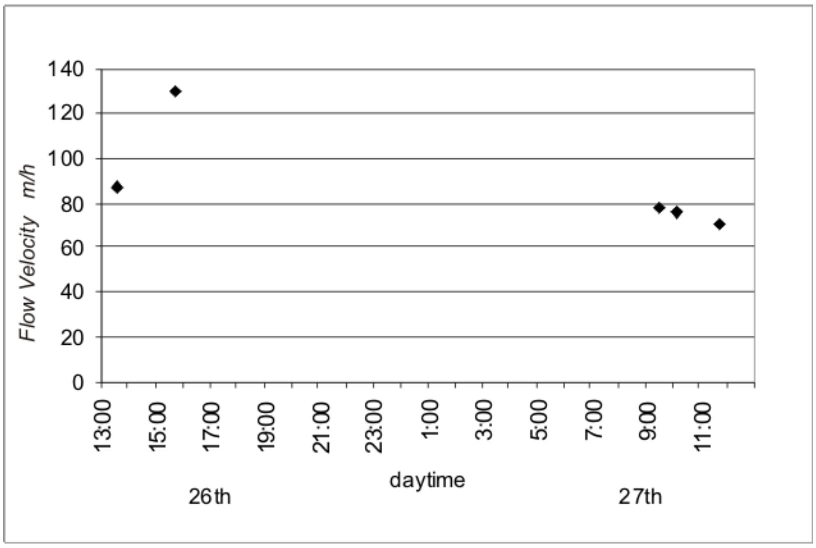


Fig. 37: Measured flow velocity of water under snow in m h^{-1} for the experimental period

6.4 Glacial Drainage System Investigation with Fluorometric Dyes

The fluorometric dye tests were done in two different measurement campaigns. It was aimed to identify different flow processes in englacial and subglacial passages during the ablation season. Measurement campaign 1 took place from the 27-28th of July whereas measurement campaign 2 was done from the 3rd to the 4th of October. The first injection sequences should point out travel times and flow path information in the high ablation season. The second test row was done at the very end of the melt season. Luckily measurement campaign 2 took place when temperatures fell suddenly under the freezing point. The temperature drop did not invert since the next spring. Therefore it was possible to observe run-off conditions suddenly influenced by a sudden stop of any melt. The temperature drop allowed a recession analysis of the receding discharge.

Previous field tests investigated water movement through snowpacks and through the basal saturated snow layer. Once meltwater reached the firn table it continues to flow downward or enters the englacial and subglacial drainage system. Meltwater can intrude the firn passages and slowly percolate through the semipermeable firn layers for then to reach the fast drainage system of englacial and subglacial channels. Notwithstanding meltwater can also be transported to the inner of the ice body relatively fast. The fragmentation of the entrance system to the englacial drainage strongly depends on local geographic conditions. It can be coarse were entrances are hardly observable (concave areas) or fine were entrances are often and usual (convex areas on the glacier). This appearance of crevasses and glacial moulins is strongly linked to local geographic conditions.

For some areas though measurement is hardly possible due to bad accessibility, no superficial meltstreams in snow or firn cover. In this field study it was difficult to sample in the nearby of the glacier fall as long as crevasses are often covered by snow and represent high risk. Moreover it was not possible to identify clearly the appearance of meltwater in this area.

The glaciated drainage area is subdivided into three basins. *Basin 1* (here) is dewatering the eastern part of the glacier involving the glacier fall and part of the glacier tongue. The other basin (here: *basin 2* includes the uppermost part of the glacier (*Oberster Boden*, *Lieslstang*) and the plane area “*Oberer Boden*” representing most of the ice masses.

Most of the following measurements show only results for *basin 2* (*Oberer Boden*). Nevertheless it was possible to estimate ranges of flow velocities.

6.4.1 Fluorometric Dye Tests

6.4.1.1 Results from Measurement Campaign 1 (Midsummer)

From the 26th to the 27th of July fluorometric dye tests were done to investigate water movement through the glacier body independent whether it flows englacially in channels or subglacial on the bedrock.
Fluormetric dye tests in these surroundings faced some difficulties regarding the needed tracer amounts as well as the injection points. In the prefield it has to be adverted that not all of the experiments did produce usable results. Lacking information about flow rates in glacial conduits resulted in too small tracer amounts for detection. Moreover generally elevated turbidity levels caused a lift of the detection limit. It was not clearly identifiable the factor with which turbidity influences dye detection.

Results from dye tracer experiments investigating the englacial and subglacial water movement on *Oberster Boden* were not representative and were not implied in this analysis.

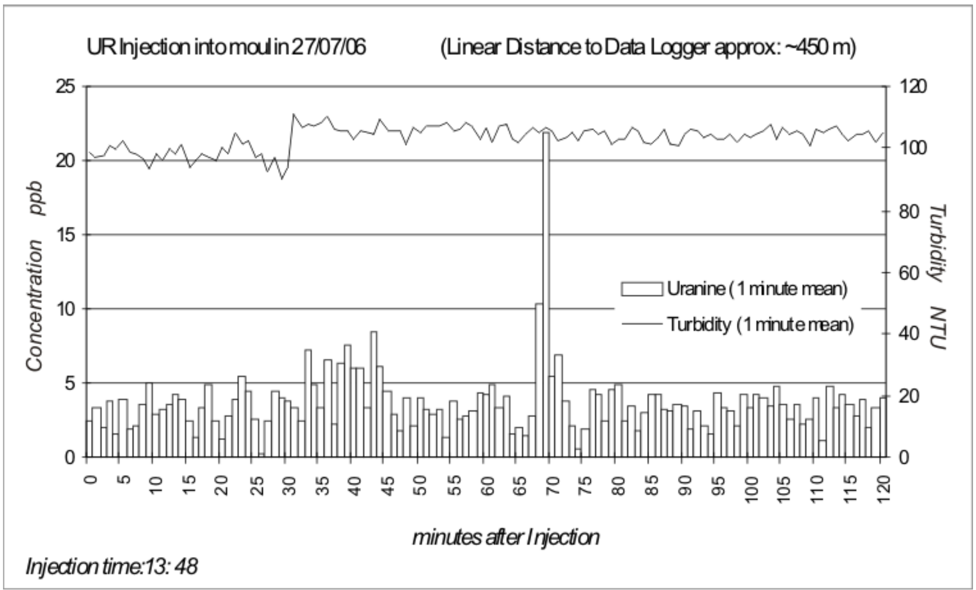


Fig. 38: Dye concentration in 1 minute means for fluorescein injection at '*Oberer Boden*' the 27th of July at 13:48. Tracer amount: 100g

The 27th of July a fluorescein injection test has been carried out on the *Oberer Boden*. 100 g of presoluted Fluorescein were dispersed directly into a supraglacial meltstream. After 30 m, the meltstream ended into a moulin. The moulin though could not discharge the complete water amounts which caused backwater effects at the entrance. The entrance to the englacial conduit system was at the edge of the ice body and entered laterally. It can be assumed that the meltstream reached bedrock as ice thicknesses were only thin in this area. Tracer dispersion in the meltstream was fast. The dye cloud moved rapidly and compact. There was no splash water since water erosion generated definite deep supraglacial channels. Approximately 3 minutes after the injection the tracer cloud disappeared under or in the ice body.

The detection device was installed in the prefield within a distance of 450 m at the outlet of the *Oberer Boden* drainage area (water fall).

The fluorometer registered turbidity (NTU) and dye tracer concentrations (ppb) in 10 seconds intervals. **Fig. 38** shows registered turbidity and dye tracer concentrations in 1 minute means. Turbidity rates were weaving around 100 NTU, with a significant jump up to approx. 110 NTU's after 30 minutes from the injection. The signal is unsteady and on a high level for the whole test period.

Fluorescein concentrations are unsteady as well and oscillate around 20 ppb. A rise of the mean dye concentrations is identifiable over the first 45 minutes. The unsteady signal makes it difficult to clearly identify increases or decreases. Nevertheless there is a first particular increase identifiable after 30 minutes, followed by a drop of the concentration signal after 40 minutes. The concentration drops to a minimum after 64 minutes for then to rise immediately to a maximum value after 69 minutes. The concentration maximum is 22.3 ppb. After 74 minutes concentrations have fallen back to a minimum. The following 45 minutes do not show particularities but weave again around 20 ppb.

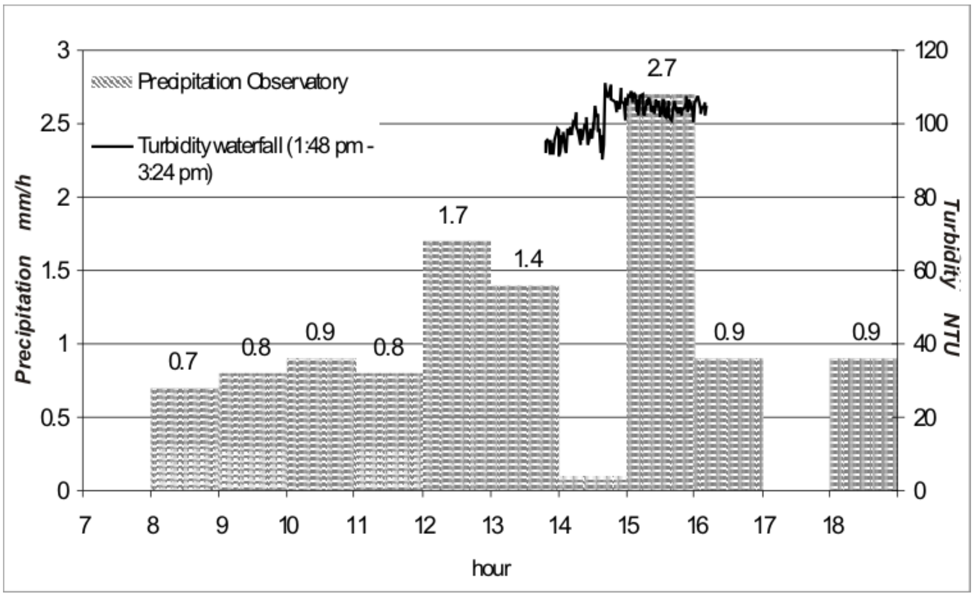


Fig. 39: Registered precipitation at the *Observatory* (*Hoher Sonnblick*) and measured turbidity for the experimental period of the 27th of July. Turbidity was registered for about 156 minutes.

The first slow increase in the concentration signal can be ascribed to the increase of turbidity over the experimental period. **Fig. 39** shows measured hourly precipitation from the meteorologic *Observatory* at the Hohe Sonnlick peak and registered turbidity at the upper outlet (water fall). The *Observatory* registered slight rainfall from 8:00 a.m. until 2:00 p.m. with the highest deposition from 12:00 a.m. to 1:00 p.m. Rainfall stopped in the following hour and started again at 3:00 p.m. in the afternoon. Measured turbidity show an oscillating but steady increase until 14:20 followed by a sudden jump. The steady increase may be caused by the passage of rainwater washing in small particles from the snow and ice surface. Slight rainfalls on the partly snow and firn covered glacier surface are slowly delivered to the drainage system. The velocity with which drainage responds to liquid deposition is therefore linked to the ability of snow or ice to store water. (see chapter 6.5.1). This capability is highly depended on snow, firn and ice characteristics as well as local topographical characteristics. Moreover it depends on the time in the annual cycle.

The turbidity increase is therefore related to the passage of rainwater deposited from 8:00 a.m. until meridiem. The sudden jump though must be related to the stronger rainfall events from 12:00 – 14:00. It is not exactly known if precipitation was deposited as a steady rainfall or a heavy short rainfall event, but personal observations point to an intensification of the rainfall over the area, around 1:00 p.m., exactly in the preparation period of the experiment. Antemeridian rainfalls where steady (drizzle).

The sudden jump in turbidity can be related to this heavier rainfall. If the rise of turbidity similar to a dye tracer breakthrough is considered, we can propose a cautious estimation of the upper drainage area response to moderate rainfalls. In this case travel times from the *Oberer Boden* area to the data logger are approximately 90-100 minutes. This is perfectly consistent to the previously measured travel times through the englacial channels after entering the moulins. Moreover particles passing in the first breakthrough at 14:20 were not those washed in from the uppermost glacierized area under the *Hoher Sonnblick* peak. Both, results from the day before (see 6.3.1.1) and later analysis will show, travel times for the farthest distant areas are significantly longer. Especially if we considered the reduced travel times in higher elevated areas due to snow cover.

After the sudden jump turbidity does not return to previous levels but slowly decreases which points to a slow effluence of the stored water.

The first hump in dye tracer concentration after 30 minutes reflects presumably the increase in turbidity and the signal override.
The steep increase of the dye concentration signal after 65 minutes from the injection though can be directly related to a tracer breakthrough. There is no further jump in possibly signal interfering turbidity observable. The breakthrough signal does not allow a delineation of proper breakthrough curve characteristics. Maximum dye tracer concentration was reached after 69 minutes.

The linear distance from the injection point to the fluorometer is 450 meters. However we must expect travel distances to be way longer since we can not even guess actual positions of en- and subglacial channels. Moreover a calculated linear travel velocity of about 0.1086 m s^{-1} resp. 391 m h^{-1} is unexpected slow. This underlines the results and estimation done in 6.1 (see also **Fig. 22**)

6.4.1.2 Results from Measurement Campaign 2 (Late summer)

A) *Oberer Boden* (3^{rd}-4^{th} of October)

Measurement campaign 2 took place in the late ablation season from the 3rd to the 4^{th} of October. Not only was it carried out to detect possible changes in the drainage behavior, but also to eliminate failures and defects in the previous midsummer campaign.

The test rows should ideally reflect dye tracer movement, flow velocities and storage in the middle and lower part of the glacierized drainage area. The drainage was sampled in three different steps covering the

a) middle part (moulin) to waterfall, resp. the movement of water under the main glacierized area the *Oberer Boden*
b) lower part from the glacier fall at the southeastern part to the glacier snout
c) lowest not glacierized area from the glacier snout to the fluorometer detection position (to register flow velocities when water has left the glacial system

No experiments were undertaken in the

a) uppermost area under the *Hoher Sonnblick* peak according to reasons described in chapter 4.2.4 and 6.1.
b) upper region of the glacier fall due to crevasses and lack of injection points and water entrances
c) water fall to glacier snout

Even if, not all main channels of the drainage area were sampled, it is yet possible to delineate simple characteristics and the storage and retention capabilities of the drainage system at the *Goldbergkees*.

Weather conditions on the test day 1 (3rd of October) were influenced by an approaching cold snap, squally winds of high velocities and drizzle. Although weather data from the *Observatory* did not show any precipitation before 11:00 a.m. light rainfall was registered at the gauge station supported by personal observations.

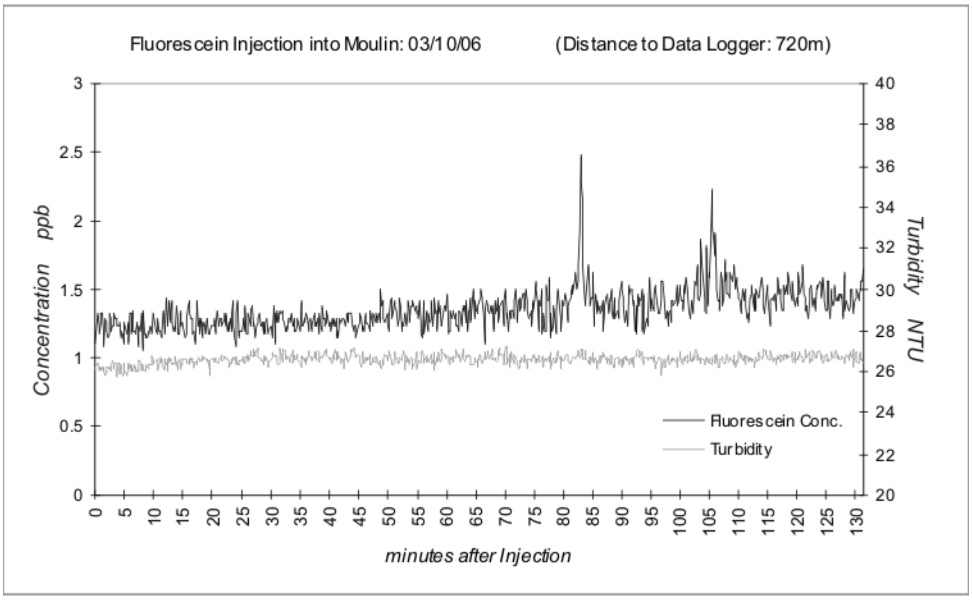


Fig. 40: Fluorescein concentration and turbidity registered in dye tracer experiment the 3rd of October. Injection point: moulin. Linear distance to fluorometer: 720 m. Injection time 12:50 a.m.

Fig. 40 shows fluorescein concentrations for dye tracer experiments carried out the 3rd of October middays. The concentration rose steadily following steady, but slow turbidity increases. After 80 minutes concentration values mounted to reach a peak (2.5 ppb) after 83 minutes from injection. After 85 minutes, concentration dropped back to normal trends. After another 15 minutes, fluorescein concentrations rose to a second hump lasting for another 5-10 minutes.

The registered fluorescein concentration lies on the lowest spectrum of detection range. Small dye amounts can be problematic if turbidity is high. Nevertheless both fluorescein risings can be assigned to the passage of dye tracer.

First dye breakthrough succeeded clearer as the following. According to this measurement the subglacial system at this time of the year must have been arborescent with a faster or slower flow resp. a shorter or longer flow path. The data does not reflect information about the position or complexity of these branchings or englacial systems.

Linear distance from the injection point to the data logger was 720 m. As pointed out in the previous chapter and in the midsummer campaign, flow paths can be considerably longer. If we consider normal flow rates in mountain torrents actual measured flow velocities seem to be relatively slow. For the distance of 720 m a total velocity of 0.145 m s^{-1} resp. 522 m h^{-1}

was calculated. It is though faster than summery measurements (0.1086 m s^{-1} for the summery tracer investigation) which could be an indication of an increase in effectiveness of the drainage system, but can not easily be compared because of different water amounts in the channels and differing sections.

A second Naphtionate dye tracer test should have lightened flow velocities in the same section but did not succeed (see **Fig. 41**)

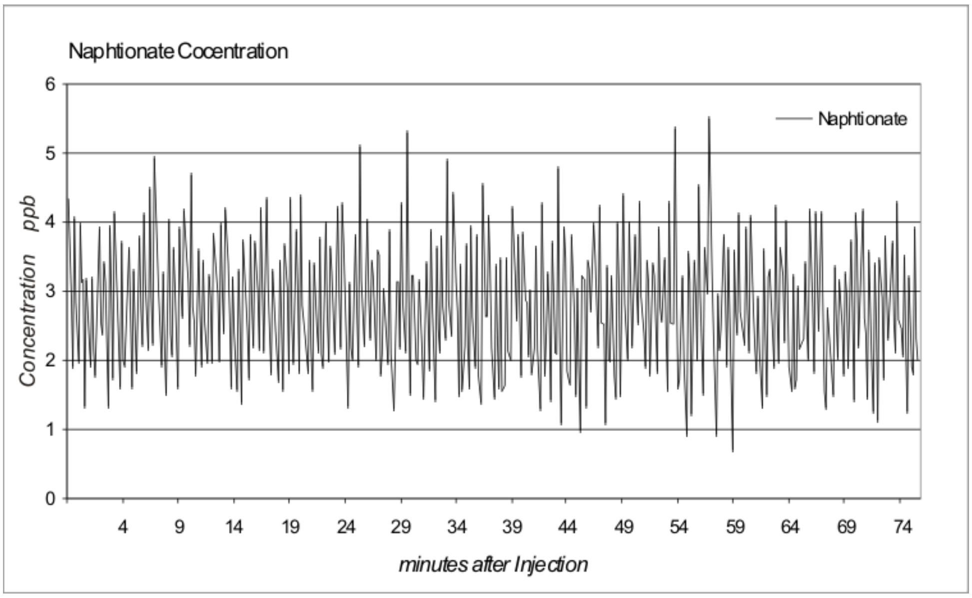
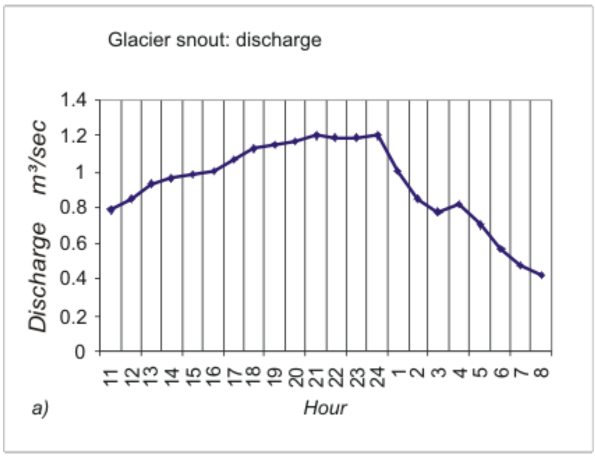


Fig. 41: Unsteady naphtionate concentration registered the 3rd of October. Injection point: moulin 1. No analysis possible

Synchronously with the first dye tracer test on the 3rd of October at the *Oberer Boden*, the fluorometer registered turbidity continuously until the next morning at 9:00 a.m. (**Fig. 42**). Run-off has an hourly data resolution whereas turbidity has a 10 seconds time resolution. Although the different time resolutions of the parameters it is possible to compare heights and lows in order to measure elapsed times water takes to flow from the water fall to the gauging station at the outlet. The read-out of the data showed similar rises and declines for the sampled period. As pointed out in chapter 4.3, rain on the ice surface or snow influences turbidity.

In this case precipitation was widespread over the whole glacier. Evidently rain washed in particles influencing the transparency of water. Peaks in turbidity are not easily assigned to discharge peaks. Nevertheless a last peak in turbidity at 3:00 a.m. can evidently be related to a final short rise in run-off at 4:00 a.m. before a continuous drop is identifiable. Moreover there is a saddle in discharge observable ending at 0:00 a.m. which is apparent in the turbidity measurement at the same time. Water takes not more than an hour to flow from the upper outlet of the *Oberer Boden* to the gauge station. The gauging station lies approximately 250 meters from the glacier snout directly on a small lake of minor depth. We can expect travel times to be significantly lower regarding the fact that water levels in lakes react with a specific retard to increased water input. Moreover we can expect water flowing fast over the cascade. It then disappears under the ice body of the glacier tongue to reach the glacier snout united with subglacial streams from the glacier fall area.



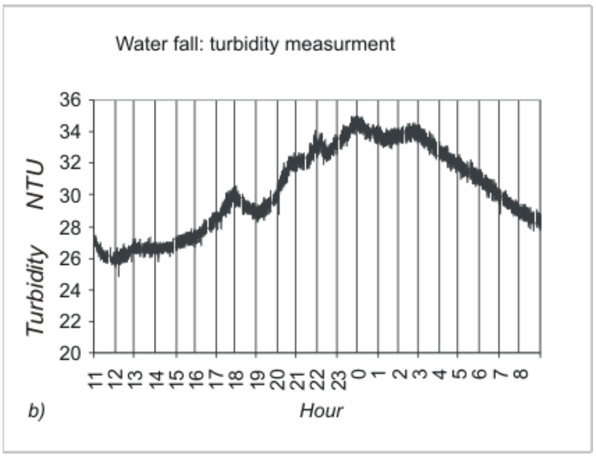


Fig. 42: a) Discharge at glacier snout from th 3rd 11:00 a.m. to 4th of October 9:00 a.m. and b) turbidity for the same time frame at water fall

B) Glacier Fall (4th of October)

In the night from the 3rd to the 4th of October, temperatures fell under the freezing point and precipitation fell solid as snow. Fresh snow amounts at 9:00 a.m. on the 4th were between 10 and 20 cm at approx. 2700 m depending on wind drift. Snow fall began approximately around 2-3:00 in the morning as temperatures have fallen under the freezing point .The sky was completely covered over the whole day.

Near the glacier fall at 2530 m a small branch of the subglacial torrent quits the ice body to flow approximately 40 m on bare rock and enters again the subglacial drainage system. At this point 50g of naphtionate and 10 g of fluorescein were released into the channel. First injection took place at 12:00 a.m., the second at 12:50 a.m. The injection point was problematic, as the torrent was turbulent, risk of splash water dispersing dye on the surrounding rocks, was great. Dispersion though should have been very fast.
The fluorometer was preinstalled at a distance of 70 m from the glacier snout.

The analysis of the breakthrough curves show an unsteady, but not rising signal for the naphtionate concentration in the first 21 minutes after the injection (**Fig. 43**). A sudden and steep increase of the concentration signal is followed by a slower decrease of the signal forming a typical tracer breakthrough curve even if superimposed by a signal noise. The tail is therefore not easily distinguishable. The average mean velocity for the linear distance of 460 m to the data logger is 0.31 m s^{-1} resp. 1116 m h^{-1}.

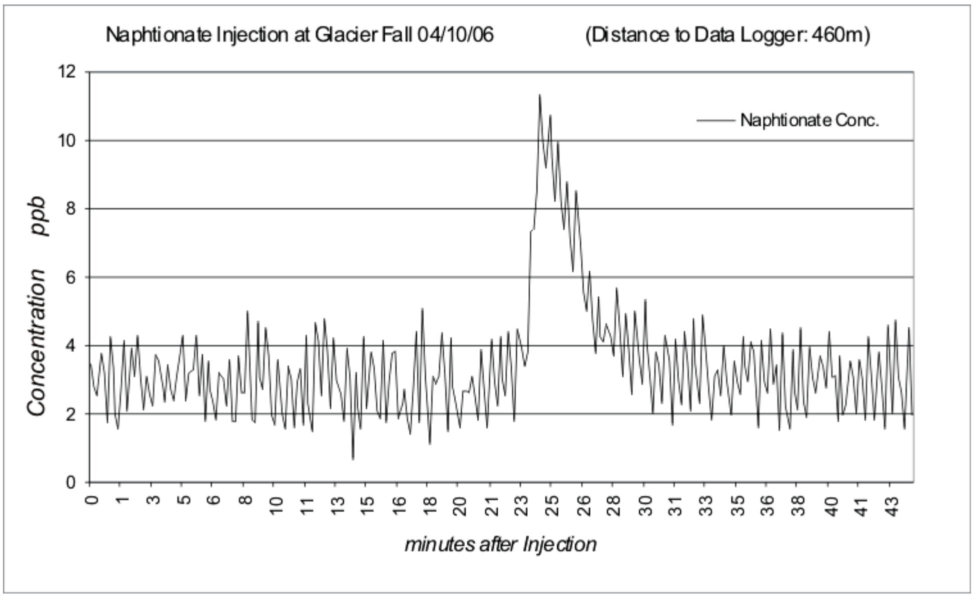


Fig. 43: Naphtionate concentration in ppb, registered at in dye test the the 4th of October. Injection point: glacier fall. Injection time: 12:00. Dye tracer amount: 50 g. Linear distance to fluorometer: 460 m.

The analyses of the second tracer test for the same injection point (**Fig. 44**) resulted in similar results as for **Fig. 43**. Peak concentration was reached after 24 minutes from the injection.

The tracer breakthrough is very clear but does not have a visible tail. It may be due to calibration or data logger postprocessing that low concentrations are not noticeable any more. Notwithstanding there are 3 smaller amplitudes after 29, 61 and 63 minutes. Turbidity did not show any variation. It could be interpreted again as an arborescent subglacial drainage system.
The calculated linear flow velocity for this dye test is 0.33 m s^{-1} resp. 1188 m h^{-1}.

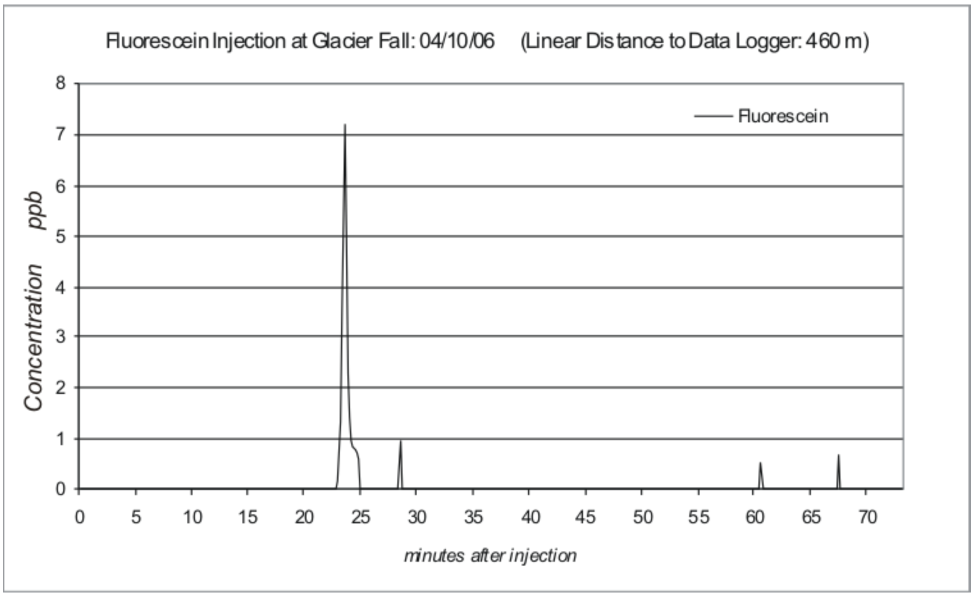


Fig. 44: Fluorescein concentration in ppb, registered in dye test the 4th of October. Injection point: glacier fall. Injection time: 12:50. Injection amount: 10 g. Linear distance to fluorometer: 460 m.

Even if the drainage system is arborescent in the tongue area higher velocities in this part are yet an indication of a general faster run-off. En- and subglacial channels have presumably been broadened by high discharge rates. In other words subglacial drainage has become efficient in the behind ablation season. Admittedly vertical heights from injection point to data logger are bigger than on *Oberer Boden*, which means inclination is higher.

C) Glacier Snout (4th of October)

Due to the fact that the outlet at the glacier snout was arborescent as well, data loggers had to be installed at 70 m from the snout where branches have unified again. The last dye tracer test has been carried out to get a fast approximation of the travel time of water between glacier snout and the fluorometer. In this case a flow velocity of 0.35 m s^{-1} resp. 1260 m h^{-1} was measured.

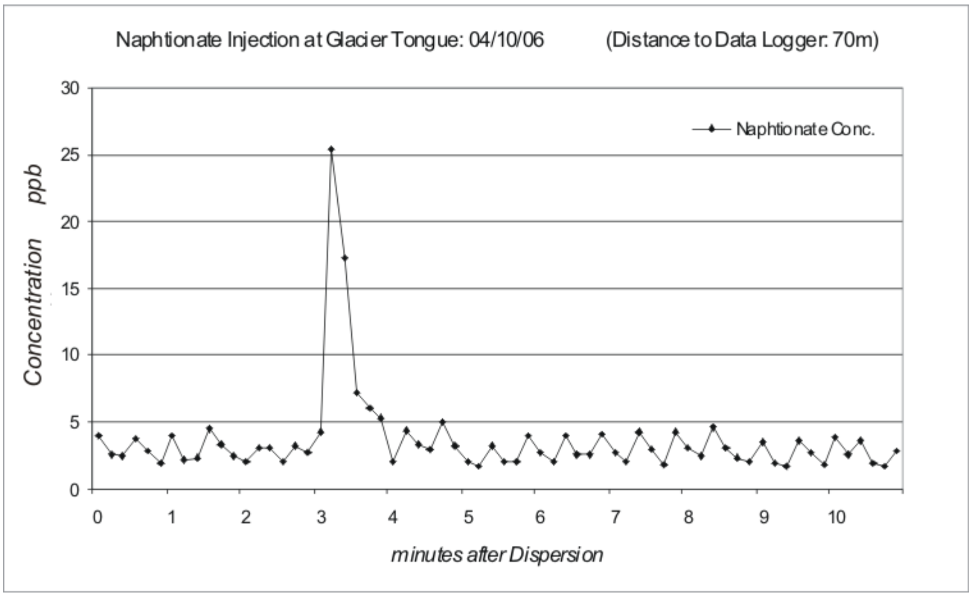


Fig. 45: Naphtionate concentration in ppb for dye tracer test at glacier snout to measure travel times from glacier snout to fluorometer position.

6.5 Results from Precipitation /Melt / Discharge Analysis

6.5.1 Results from Precipitation / Discharge Analysis

Dye tracer tests for investigating storage capabilities as well as transit times of water in a glacierized surrounding can be very useful. Nevertheless they face a series of difficulties mentioned in previous chapters. Especially for complex drainage areas tracer measurements are time consuming and elaborate. Due to their subdivisions and local varying characteristics the whole system has to be investigated which results in a high effort.

The excellent data base for the investigation area, allows though further investigation by analyzing the interrelation of meteorologic conditions with discharge data.
Melt and rainfall causes variances in the discharge amounts. High amplitudes in discharge and exceptional meteorological events can be connected.

Therefore discharge analysis of summery rainfall events where done to gather further information about retardation effects and storage effects of glaciers.
In a first step, all days on which precipitation was registered, were separated from dry days. The rainfall events were classified after their intensity and a possible connection to an abnormal discharge hydrograph. Only on few days a clear response of the hydrograph was observable. Especially very unproductive rainfalls are hardly separable from normal discharge hydrographs or show no clear breakthrough curve and are simply superimposed on the normal hydrograph.
Moreover precipitation data from the *Hoher Sonnblick Observatory* showed smaller and slower impact in discharge and is difficult to separate from precipitation at the tongue.

Though we can assume that widespread precipitation is registered on both precipitation measurements stations, whereas local precipitation or thunderstorms are registered either on the *Observatory* or at the tongue. Precipitation at the *Observatory* can be very local and therefore embrace only the peak area and therefore the uppermost glacierized area. Rainfalls registered at the tongue are more widespread and embrace oftenly the *Oberer Boden* without being registered at the *Observatory*. In fact correlation matrices in Tab. 7, reflect almost no interrelation between the deposition at the glacier tongue and the highest areas.
In a coarser comparison of rainfall on rainy days there are though similarities. Rainfall, if registered for both locations can be displaced in time and amount. Meteorologic conditions in high mountain areas are very inhomogeneous. Especially precipitation can be locally

limited and often depend on special topographic characteristics. Hence results in among calculated correlations.

Tab. 7: Correlation between Precipitation at the *Observatory* (PRECIP_OB) and discharge gauge (PRECI_TO) computed with SPSS 11.5 showing no similarities in time and precipitation quantities.

		PRECI_OB	PRECI_TO
PRECI_OB	Pearson Correlation	1	.097(**)
	Sig. (2-tailed)	.	.000
	N	2771	2771
PRECI_TO	Pearson Correlation	**.097(**)**	1
	Sig. (2-tailed)	.000	.
	N	2771	2771

** **Correlation is significant at the 0.01 level** (2-tailed).

6.5.1.1 Rainfall on Lower Glacier Area

Summery rainfall in mountainous areas can be widespread, normally with lesser intensities caused by rain fronts or local with higher intensities, caused by convective thunderstorms. Highest intensities in rainfall were registered mostly at the end of July and in August. However rainfall events in August did not show clear amplitudes in the discharge hydrograph and could not be used in this analysis. Moreover the gauge station had 2 blackouts affecting exactly 2 main rainfall events. Therefore only few adequate hydrographs remained to be integrated into this comparison.

Exemplary two resp. three main events are described here. Both events happened at the end of July at the 26th, 28th and 29th (**Fig. 46**). The 26th of July has been used as a field investigation day of this study to draw movement of water on the firn table. The figure shows discharge amplitudes and registered precipitation at the gauging station. It illustrates on first sight the impact of heavy rainfall on the run-off. Whereas the 25th of July shows a normal hydrograph with a saddle and a receding limp, 26th mirrors sudden amplitude in discharge. Hydrographs from 29th and 30th of July show similar reactions.

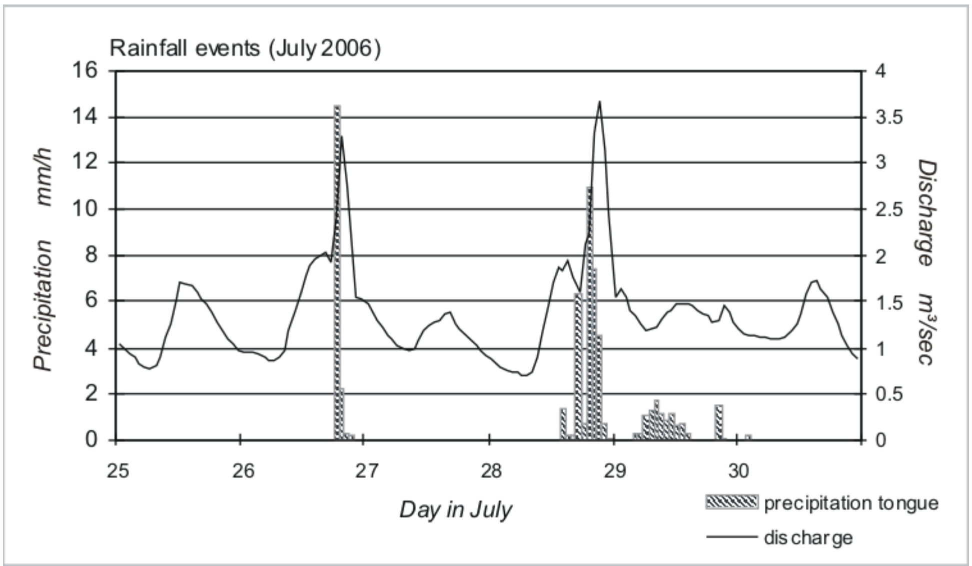


Fig. 46: Rainfall on lower glacierized area resp. *Oberer Boden* and glacier tongue. Discharge hydrograph variance for "rainy" and "dry" days. Amplitudes are directly related to precipitation.

Fig. 47 reflects discharge, global radiation (irradiance) and precipitation measurements on the 26^{th} of July. Noticeable are the dry and sunny conditions during daytime with high radiation. Conditions evoked high melt rates and a typical delayed increase in discharge at the gauge station. In the early evening weather conditions changed rapidly as a thunderstorm with high rain intensities passed the glacier area. These facts can be underlined by personal observations. From 19:00 p.m. to 20:00 p.m. 14.6 mm of water has been registered at the glacier tongue whereas the thunderstorm only grazed the uppermost areas. Most of the precipitation must have fallen on the wider area of *Oberer Boden* and at the glacier tongue. The discharge hydrograph reacted intensively to the rainfall showing a superimposed peak on the normal melt induced hydrograph. The peak discharge is 75 % higher than the maximum melt related discharge of this day and occurs only 1 hour after the deposition.

The steep increase in run-off is followed by a short peak and a fast fall. The breakthrough curve tail assigned to rainfall is not noticeable because it is superimposed the normal receding limp of the hydrograph. It could be separated mathematically but is rather complicated because of the inhomogeneousity of daily glacier hydrographs.

The deposited water amounts near the discharge gauge contribute immediately to an increase in discharge. Though, the highest discharge amount is registered after a lag-time of 1 hour which represents the mean travel time of water from the *Oberer Boden* to the snout. Certainly flow velocities are changing with the run-off amount. Nevertheless it can give important evidence for the short term water storage of the bare-ice areas at the *Oberer Boden* and the tongue. The reaction of run-off to precipitation is highly depending on storage characteristics of snow, firn and ice. Roughly it is though to say, that snow will show better storage capabilities than bare ice (firn, glacier ice). In other words discharge from lower areas on the glacier will happen effectively and fast because of the mostly missing snow overlay. Storage of water in the ablation zone be it meltwater or rainwater, is of short term and could be compared with bare rock once channels are wide enough to swallow transported water amounts.

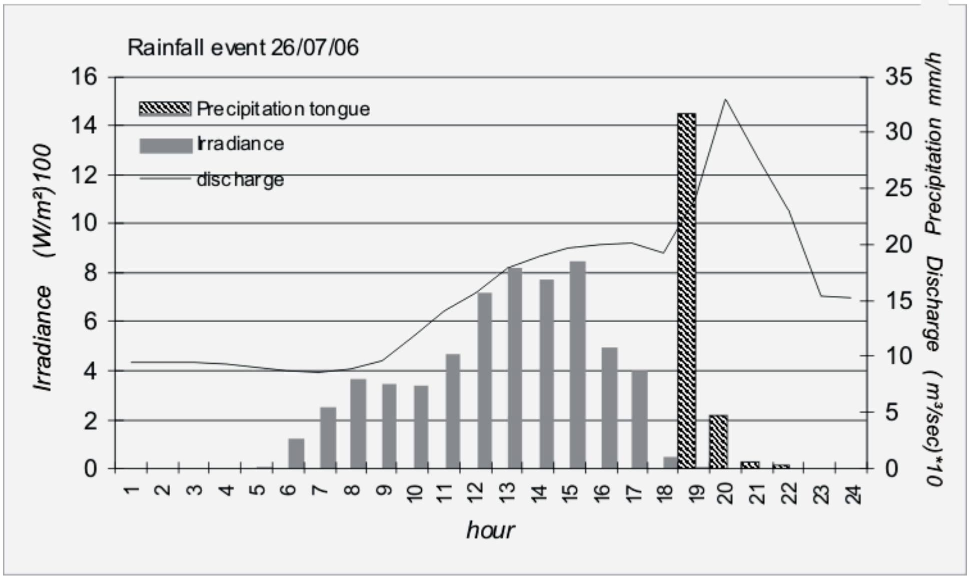


Fig. 47: Rainfall event on the 26th of July registered at gauge station. Irradiance and discharge for the given timeframe

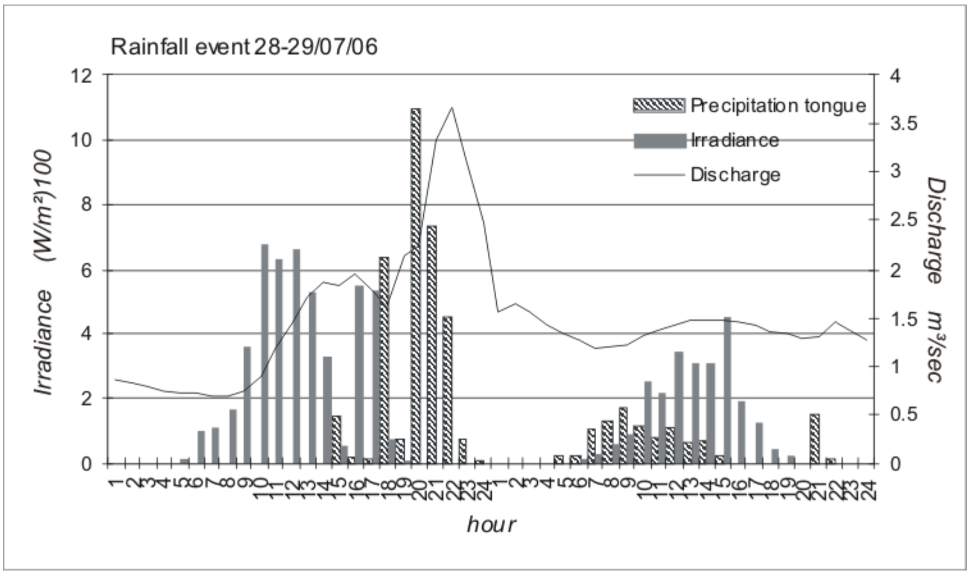


Fig. 48: Rainfall event on the 28th and 29th of July registered at gauge station. Irradiance and discharge for given time frame

The rainfall event on the 28th of July shows similar characteristics (**Fig. 48**). The melt related normal discharge hydrograph is superimposed by a followed summery thunderstorm. In this case though precipitation amounts were greater and rainfall was lasting longer. Nevertheless travel times were similar. Both daytime rainfalls at 15:00 p.m. and vespertine rainfalls at 18:00 p.m. and 20:00 can be related to an increase at the gauge station approx. 1 hour later.

6.5.1.2 Rainfall on Upper Glacier Area

Discharge hydrograph is reacting differently to precipitation in the upper glacierized area. Not only is it characterized by higher distances to the gauging station but does hold different physical characteristics than lower parts of the glacier as well. Roughly we can say that characteristics change under and above the equilibrium line. The accumulation zone of a glacier is characterized by greater snow amounts over the total annual cycle but also can it endure the ablation period or at least last longer. This is why storage properties of the upper glacierized zones are more effective.

Though, it is difficult to separate rainfall events which occurred only on the uppermost areas. The rainfall event shown in **Fig. 49** is a perfect example of a local rainfall on the uppermost glacierized area resp. the so called *Oberster Boden* or *Lieslstang*. The discharge hydrograph shows an increase of run-off during the night and reaches a peak at 8:00 a.m. in the morning. Afterwards run-off is decreasing by approx. 10% of the maximum value over the next 4 hours. At 13:00 p.m. run-off is surging again to reach a peak at 16:00 a.m. in the afternoon. It reduces in the following hours by half the maximum value.

Precipitation was scarcely registered at the glacier tongue but showed high values at the *Observatory* pointing to a local precipitation event near the *Hoher Sonnblick* peak. Nocturnal rainfalls remained under 0.5 mm h^{-1}. At 9:00 in the morning a heavy rainfall was registered depositing nearly 18 mm in one hour. After 4 hours first light indices are found that rain waters reached the gauge station. 7 hours after the rainfall discharge reached a peak and dropped relatively fast after that. The hydrograph clearly shows the passage of the rainwater at the gauging station. Nor was it radiation related meltwater (covered sky, rainy conditions) or precipitation on the lower glacier area. First contact with rain waters from the upper part can be detected after 4 hours, whereas the peak discharge is reached after 6-7 hours. It perfectly fits to the fact that longer flow distances stretch the base of the superimposed breakthrough curve. Rainfall near the gauge would cause a compact base.

At the receding limb of the discharge hydrograph at 0:00 on the 19th of August there is a further sign of a rainwater breakthrough. Although there was no proper increase recorded there is still a temporary slowing of the normal decrease in discharge amount observable. This hump can be assigned to a weak rainfall event occurring 8-12 hours before. Lag time between precipitation and run-off is therefore significantly longer than in the previous analysis. It is an argument for a slower reaction of discharge hydrograph for weak rainfall events. The reasoned deduction is that heavy rainfall can fill storage compounds in snow, firn and ice faster than weak rainfall. If the short term storage capabilities are overstepped the drainage system must conduct the water to drainage system. Moreover it will produce higher pressures and increase flow velocities in granular veins, supra -, en- and subglacial channels. Minor water input into the glacial drainage system through light rainfall fills the storage bodies rather slowly. Latter will conduct the rainwater slowly to the effective drainage system.

A quantification of the discharge retardation for variable intense rainfall is not possible due to small data quantities and the persistently changing drainage characteristics during the ablation season. The example though shows evidence for a decrease in lag time for heavy rainfall.

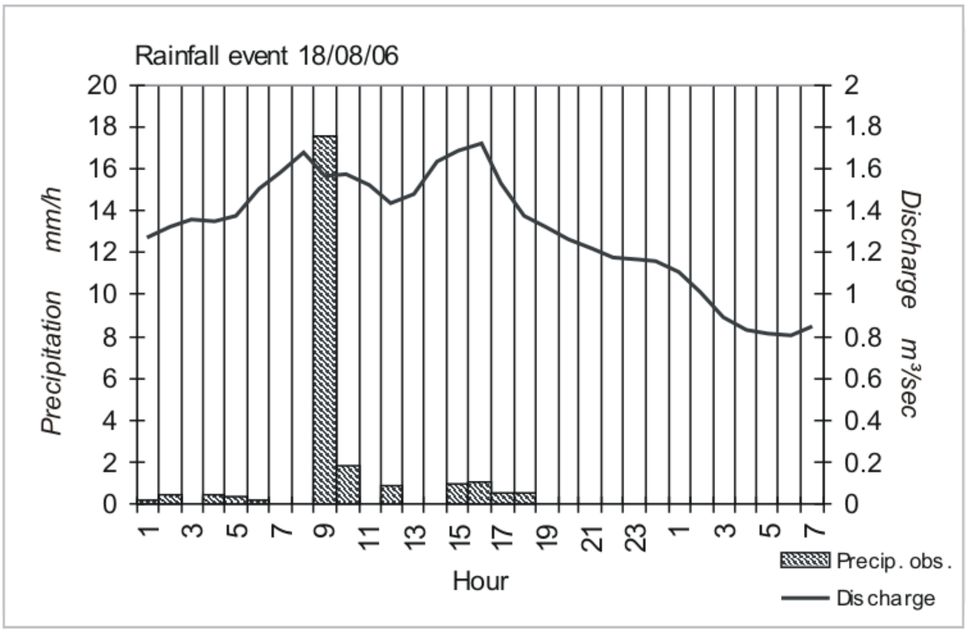


Fig. 49: Rainfall event on the 18th of August registered at the *Hoher Sonnblick Observatory* and discharge at glacier tongue.

6.5.2 Results from Dry Melt Day Analysis

6.5.2.1 Interaction between Irradiance, Temperature and Melt

The dominating meteorologic factor causing snow or glacial melt is radiation. Temperature plays only a subordinate role. The cold surface generates a relatively stable atmospheric layer with almost no turbulences. Therefore heat fluxes induced by turbulences play only a minor role even if it is highly dependent on meteorologic conditions (Lang et al., 1977). Solid and organic matter pollutes the glacier surface and leads to a decrease of albedo. The darker particles on the surface favour a spatially limited warming. Energy transfer to the ice surface causes melt. The higher the albedo the smaller is the influence of irradiance regarding melt. Melt on fresh fallen “clean” snow is evoked stronger by temperature than by radiation.

Radiation is the meteorologic forcing for temperature to increase. Especially on dry melt days with high radiation, temperature increase is directly linked to incoming global radiation. Temperature is therefore causing melt predominantly as a consequence of radiation.

As shown in **Fig. 50**, temperature is reacting with retardation to irradiance variations. The temperature increase in the early morning hours is fast and chronologically highly linked to the increase of irradiance. Small deviations can be evoked by air mass movements in the early morning hours. Irradiance reaches a saddle around midday. The temperature reaction to the increase of irradiance in the forenoon is not completely explicable by incoming global radiation. Highest mean temperatures were reached at 15:00 to 16:00 in the afternoon. Temperature reacts therefore with approximately 3 hours of retardation. The decrease of temperature in the late afternoon and evening is though similar to the recession of global radiation until sundown.

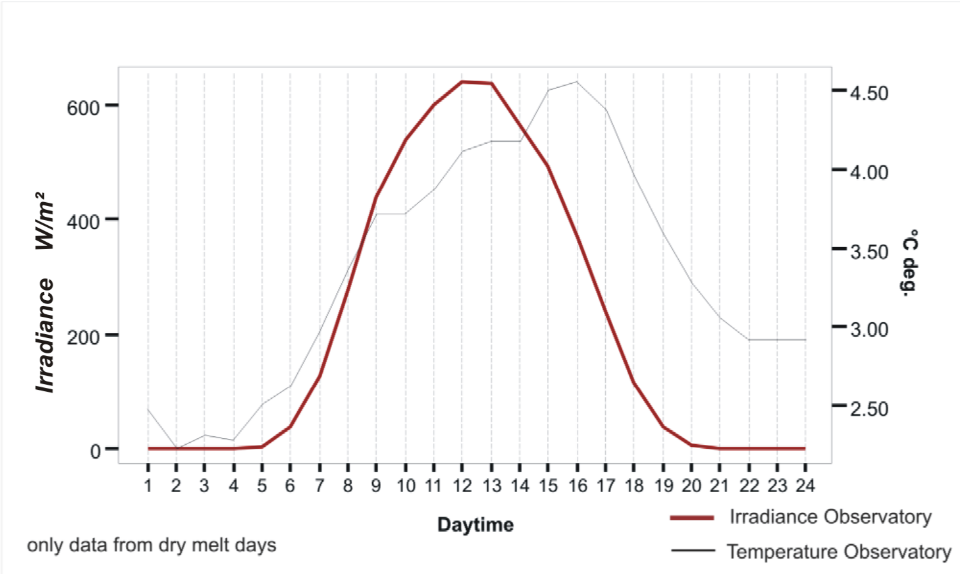


Fig. 50: Total mean hourly Irradiance and temperature at *Observatory* for dry melt days in the ablation season 2006

As a consequence to the relation of global radiation and temperature, run-off must be correlating with measured temperatures at the meteorologic stations. In fact correlation matrices show a relatively strong connection to the measured discharge even if runoff values are affected by scattering. Though, highest run-off values are related to rainfall events (**Fig. 51**). Moreover the chronologically retarded discharge leads to not very clear correlations.

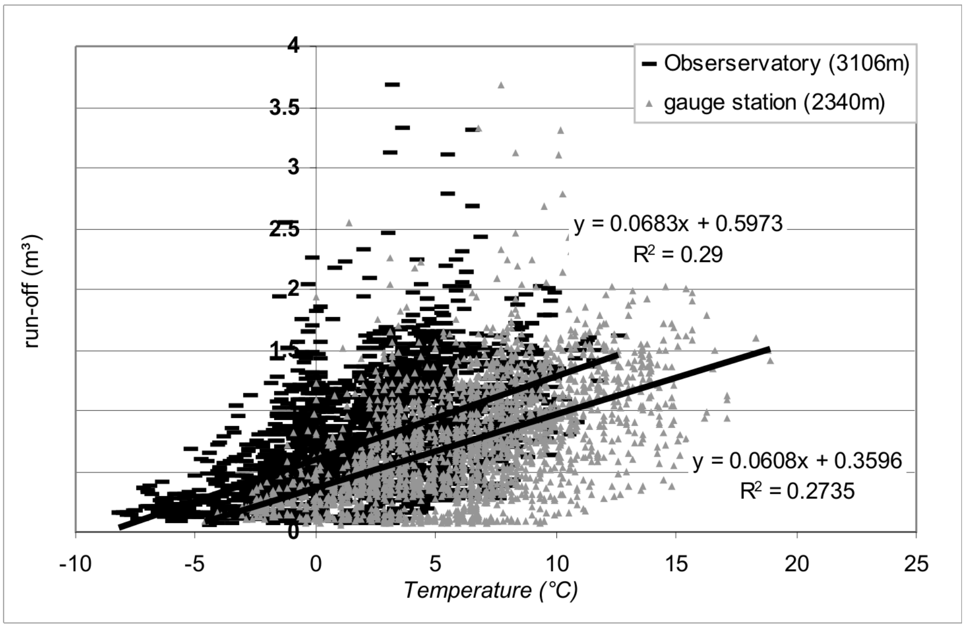


Fig. 51: Correlation between run-off and temperature at the *Observatory* (*Hoher Sonnblick*) and the gauge station (glacier tongue)

The lag time between melt and run-off (considering temperature as the driving meteorologic forcing for once), is lesser than for irradiance and run-off. Moreover irradiance is dropping to zero by night causing melt to stop at the glacier surface. Discharge does not drop to zero because of the retarded run-off and slowly emptying storage bodies. Correlation is therefore hardly visible as long as retardation of runoff can cause a peak discharge at 20:00 p.m., at a time were almost no melt is registered any more.

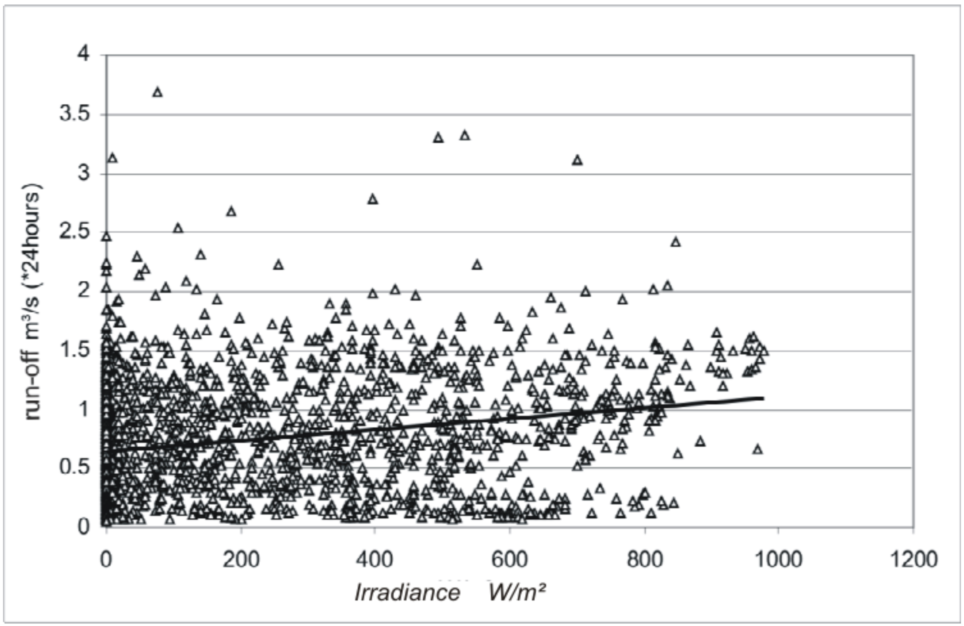


Fig. 52: Correlation between run-off at glacier tongue and irradiance at the *Observatory* in W / m².

6.5.2.2 Discharge analysis for dry melt days

Discharge analyses for the ablation season provide additional information regarding the lag time between melt and run-off as well as water storage abilities in general. Without the application of dye tracers it is possible to delineate discharge and retardation trends in the ablation season of 2006. The information output from discharge analyses, provide additional information backing dye tracer experiment results.

For an evaluable analyze though, days on which rainfall occurred had to be separated from so-called "dry melt days". Rainfall can occur at any time of the day and therefore distort melt related hydrograph characteristics. Dry melt days are characterized by no registered wet deposition and high melt rates, related to high incoming global radiation.
For the investigated period only few proper "melt days" were found (n= 30). 9, 14, 7 days in July, September and October were identified as proper dry melt days (**Fig. 53**). Meteorologic conditions in August were changing and wet, so that no dry melt days could be identified.

A first graphical analyze displaying irradiance and discharge shows a retardation of discharge on all days. Both irradiance and discharge show different general levels for selected periods in the ablation season. Irradiance is slowly decreasing over the ablation season as the irradiation angle is decreasing and days are getting shorter. Consequently mean daily discharge should decrease with decreasing melt rates.

Such a trend could be observed for the complete ablation period but was not provable for the selected dry melt days. On the contrary, a series of sequent dry melt days resulted in an increase of mean discharge amounts even if irradiance is decreasing from day to day (**Fig. 54** and **Fig. 55**).

Estimated discharge levels for dry melt days in July oscillated approx. around 1 $m^3 s^{-1}$ and reached a maximum the 14th of July. Discharge levels in September were around 0.8 in the first half and 0.7 $m^3 s^{-1}$ in the second half.

At the beginning of October a cold snap with snowfall caused an irruption in melt. In this case temperatures under the freezing point were mostly responsible for low melt rates. Decreased radiation and shorter sunshine duration favored low discharge rates as well. The discharge rates did not rebound any more for the ending ablation season. Discharge rates were only around 0.1 $m^3 s^{-1}$ for dry melt days in October (**Fig. 53**, d)).

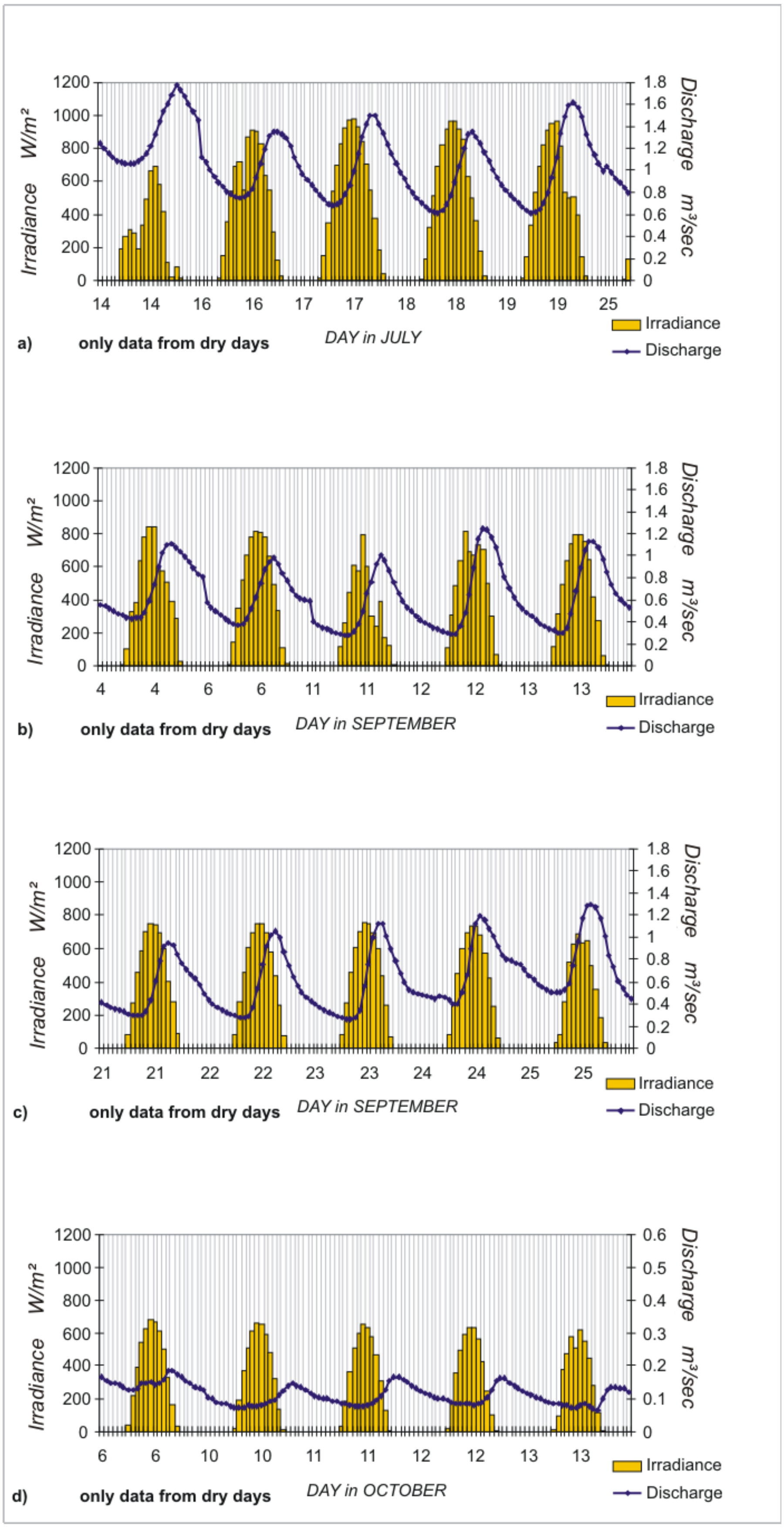


Fig. 53: Irradiance and discharge for selected dry melt days in the ablation period 2006. August not present since no dry melt days were registered.

Although irradiance is continuously decreasing over the ablation period which should result in a decrease of discharge rates, the opposite was observed for the period from the 21st to the 25th of September. Conditions in the prefield were rather cold with unproductive rainfalls. Discharge oscillated around 0.5 $m^3 s^{-1}$ at the basins outlet. Nevertheless a steady increase in run-off is registered. The situation can be explained by the storage capacities of a glacier. Even if the discharge system is rather effective at the end of the ablation season it still holds back water and conducts it slowly to the run-off. According to snow and ice behavior at low water pressures in the snow, firn and ice body, the retardation is higher for lesser free water quantities. Pores, rinulets and channels can hold water longer if the discharge system is not saturated. In this case drainage systems are expected not to be saturated. The slow outflow representing the breakthrough hydrographs tail, underlies the increasing discharge of the following day. The signal of the following day is therefore superimposed to the base flow and the decreasing run-off produced the previous day. If outflow is stretched and near the base flow-level it produces a trend like increase in mean discharge if the following days are affected by similar meteorologic conditions. **Fig. 64** shows a steady increase in mean daily discharge amounts.

This development though was not observable for dry melt days in July with similar meteorologic conditions. The steady increase in melt related discharge though, will slow down and stop as certain discharge levels are reached. According to the discharge data of 2006 and the special characteristics of the drainage system for the examined period, this discharge level must be reached somewhere around 0.8 and 1.1 $m^3 s^{-1}$ as long as no increase was observed in the "dry melt day"- phase of July (**Fig. 62**, a)).

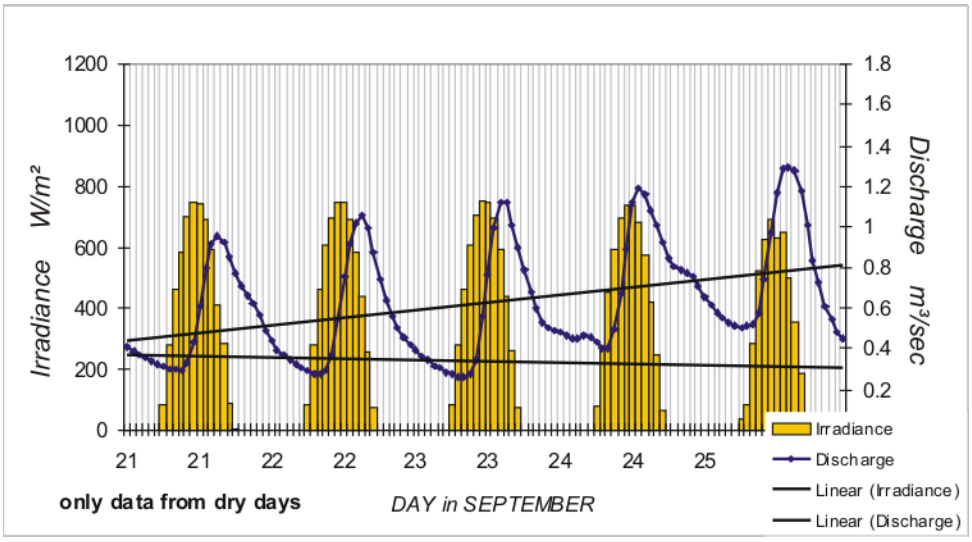


Fig. 54: Irradiance and discharge for selected days in September 2006. The black lines re-presents the linear trend function for mean discharge in the selected period and the linear trend function for irradiance

The steady decrease in total incoming radiation over the summery ablation season is displayed in **Fig. 55**. For this visualization only dry melt days were implicated. August is not present which explains also the sudden drop at n=8 (x-axis). Irradiance in July is registered already between 4:00-5:00 a.m. and lasted until 19:00-20:00 p.m. showing a typical bell form. Sun duration in September and October was 2-4 hours resp. 4 hours. On clear days in July with almost no cloud cover an energy input of 900 W m^{-2} are attained.

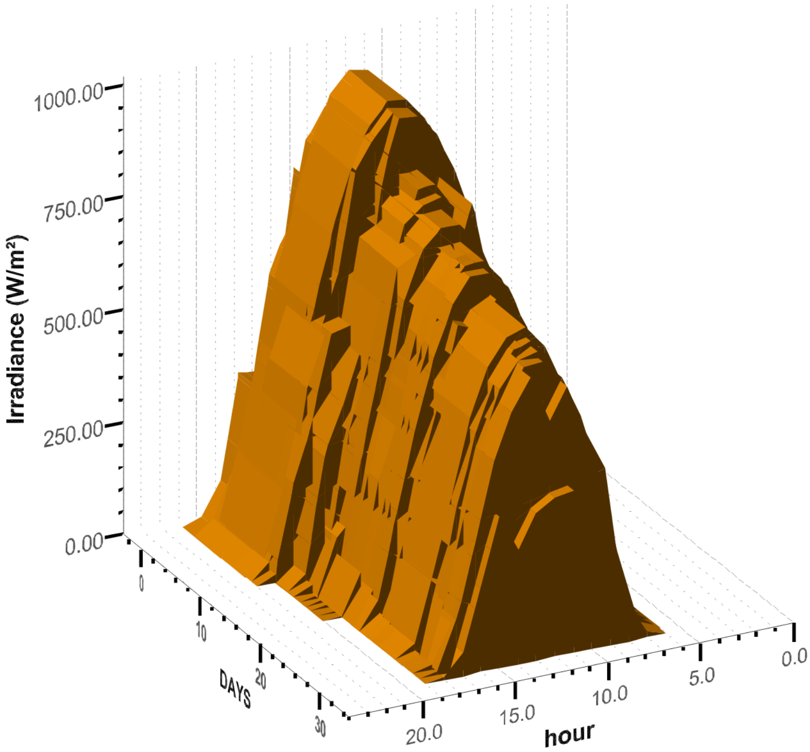


Fig. 55: Irradiance for selected dry melt days in the ablation season 2006 showing July at the back, September in the middle and October at the front. August is not embraced in this figure as it was no complete dry day registered. (July N= 9, September N=14, October N=7)

The retardation of discharge is caused by two main factors:

1. flow path resp. distance from melt location to water gauge
2. effectiveness of the drainage system resp. the storage capacities of the snow, firn, ice bodies

Most of the retardation effect of run-off in the basin is caused simply by the distance, water has to travel until it reaches the outlet at the glacier snout. The distance varies between only a few hundred meters to more than 5 kilometers for the uppermost snow or ice covered areas. Moreover we can assume higher melt on areas with a favorable inclination. Shadowing effects will avoid melt in the earliest morning hours and reduce it in the late afternoon hours, even if registered sun duration is longer for the peak area. We can assume that waters, registered in the peak discharge period, to arise mainly from the *Oberer Boden*, and higher areas. According to results from dye tracer experiments, melt water produced at midday at the tongue will have shorter travel times to the gauge. The hydrographs tail though, is assumed to arise from the steadily diminishing outflow of all snow and ice bodies of the glacierized area, if radiation induced melt has stopped.

The effectiveness of the drainage systems on glaciers is changing and reaches a maximum at the end of the ablation season. Veins, rinulets and channels in snow, firn and ice grow in diameter because of pore pressure variations and melt on the contact surface of meltwater flows and ice. Transport capacities are rising with increasing cross- sectional areas. Moreover backwater effects are less likely if transport capacities are increased so that storage is diminished.

Backwater effects in the drainage of the *Goldbergkees* are of lesser importance for the dry melt phases in July, August and September than they are in the initial ablation season. The scarcely developed drainage system at the beginning of the melt period is mostly a relict of the previous year. Whereas we can expect small channels to be sealed by freezing in the wintery period, channels with large diameters endure the accumulation period and are reactivated as soon as melt begins anew.

Nevertheless we do not know much about wintery channel characteristics and can only estimate possible characteristics. Furthermore, it was not possible to investigate drainage behavior in the beginning melt phase as discharge gauge was online only from the beginning of July. Storage capacities in June should be of great interest as long as we can suspect the drainage system to react differently to increased run-off than in the high ablation season.

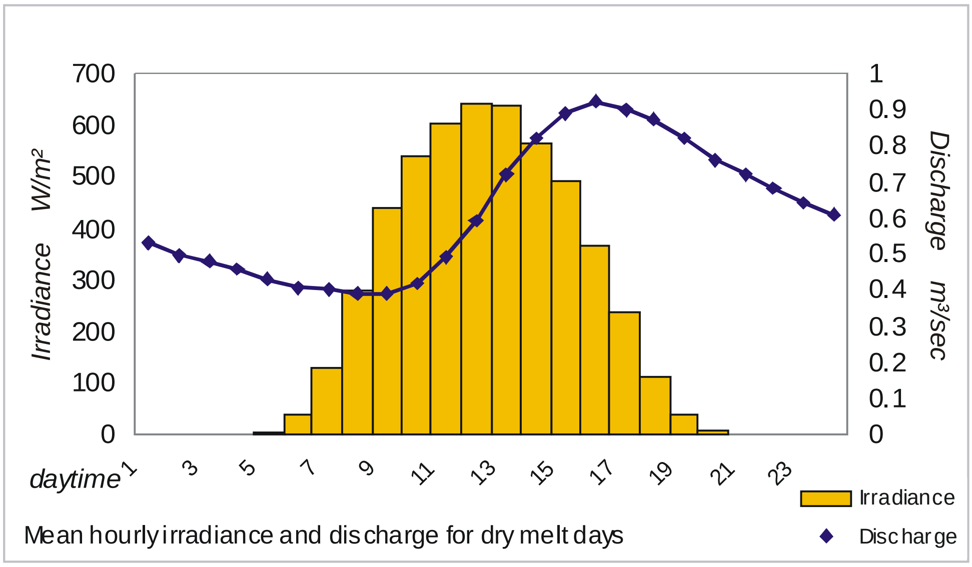


Fig. 56: Mean hourly radiation and discharge amounts for the complete ablation season 2006 for dry melt days.

Retardation analysis embracing all dry melt days in the ablation season, show the peak energy input at 12:00 to 13:00 and a peak discharge at 16:00 – 17:00 p.m. This results in a mean discharge delay of approximately 4-5 hours (**Fig. 56**). According to dye tracer experiment results (chapter 6.4), travel times from the main areas on *Oberer Boden* and *Lieslstang* are significantly shorter in englacial channels. Due to this reason, approx. half of the run-off delay must be assigned to a slow movement of water through the snow and firn and over the glacier surface itself. Mainly structural characteristics of the glacier surface (snow cover, firn, ice surface) lead to lower flow velocities than in en- and subglacial conduits. In contrast to a laminar percolation and movement on the surface, the channelized flow with higher water amounts in conduits favors higher flow velocities. Compared to channeled flow paths, the structural roughness of the glacier surface reduces flow velocities. The collection of meltwater on a glacier can be compared to an arborescent hydrologic system. Following the orographic flow direction branches are getting wider, transporting more meltwater with an increased velocity due to reduced roughness.

Tab. 8: Computed correlation between Irradiance and 3 -7 hours displaced Run-off for dry melt days show highest values between 4-5 hours after melt.

		7 hours	6 hours	5 hours	4 hours	3 hours	Run-off
Run-off	Pearson Correlation	**.453(**)**	**.506(**)**	**.530(**)**	**.518(**)**	**.470(**)**	1
	Sig. (2-tailed)	.000	.000	.000	.000	.000	.
	N	696	696	696	696	696	696

** Correlation is significant at the 0.01 level (2-tailed).

The mean retardation effect over the whole ablation period of about 4-5 hours is only an indication of run-off delays, but does not mirror the transformation of the drainage system over the ablation period. In order to show an evidence of increasing effectiveness regarding the drainage system, monthly dry melt days were separated to calculate hourly-monthly delays. **Fig. 57** renders hourly mean discharge hydrographs for July, September and October. It demonstrates the decrease in run-off delay. Whereas July shows a retardation of approx. 6 hours in peak discharge, September retardation is only about 4 hours. Values from October hydrographs are difficultly interpretable, since no clear peak discharge is observable. The hydrographs base though is similar to September observations.

The graphic produces crucial conclusions on the transformation of the discharge system in the ablation season 2006. Run-off in the late ablation season is significantly faster than in the first half. Unfortunately June hydrographs can not be evaluated because of missing data.

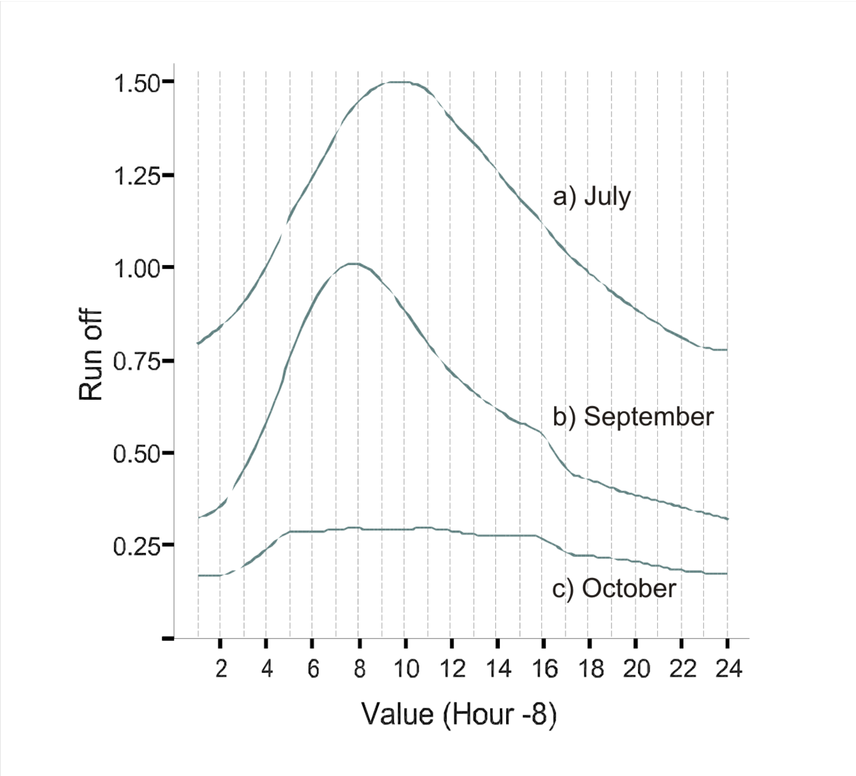


Fig. 57: Mean hourly discharge for dry melt days in a) July, b) September and c) October. Y-axis is showing mean discharge in m^3/s, X-axis hour -8 (for real hour value add 8).

The beginning ablation season must therefore be dominated by a slower flow in hardly developed glacial drainage channels. Storage capabilities are higher and are mostly relatable to the physical characteristics of the snow and firn cover.

The loss of snow in the second third of the ablation season leads to a decrease in storage capacities and to a faster transfer to the en- and subglacial system. The deglaciated areas are building arborescent and effective surface drainage channels. Moreover, sealed and lesser developed firn and glacial conduits are opened and expanded.

In a third phase the snow cover is completely lost except for accumulation areas. Firn and bare ice can only detain small amounts of water. The entrances to the englacial system grew in size and number over the ablation period.

Changes in the behavior of run-off delay happen mostly rather permanent as a slow process over the melt season, than fitfully as a sudden change.

7 Summary and Conclusion

The drainage system of the *Goldbergkees* is highly complex. The primary processes during melt or rainfall induced run-off are not easily outlined by simple field observations. Further investigations by dye tracer sampling and discharge hydrograph analysis were believed to support initially established assumptions.

Nevertheless the conclusions to withdraw from analysis of observed and computed drainage were important for further investigations in this study. The field observations in the run-up and during the whole experimental process could partly record the development of superficial drainage channels.

Generally, the drainage basin is highly fragmented and subdivided into two main drainage areas, dividing the glaciated area of *Goldbergkees*. This fact complicated investigations considerable. Neither allows it the uniform handling of discharge data from rainfall or melt, nor is it possible to evaluate tracer experiments for the whole area. Moreover, a further subdivision caused by a deglaciated part between *Oberster Boden* and *Oberer Boden*, complicated analysis as well.
The ice-free part showing a steep rock face is characterised by an huge number of small sized discharge pathways dewatering the uppermost part of the glacier. The drainage in this part of the glacier is hardly channeled. The recording of these small sized channels show the fragmentation of run-off from *Oberster Boden*. Run-off from the main glacierized area *Oberer Boden* is mostly en,- respectively subglacial. Even if melt and rainwater is collected superficially in gullies it suddenly can disappear and enter the ice body. The area of *Oberer Boden* is characterized by two glacier moulins and a crevasse area on the southern edge near the glacier fall. Both glacier moulins, collecting waters on the plane area of *Oberer Boden* and crevasses on the steeper areas with higher ice flow velocity rates, represent the main entrances to the en- and subglacial drainage system. Observations in high ablation season showed widespread superficial movement of water on the *Oberer Boden* but not on the *Oberster Boden*. This observation must be linked to the surface structure of the glacier, showing blank ice on the lower and firn or snow on the upper glacier areas. Run-off in the tongue area is mostly subglacial. The two main run-off branches from the main basins reunion in the very last part, directly under the glacier tongue for then to abandon the glacier drainage system at the glacier snout.

En- and subglacial channels are not derivable from simple observations. Not only do we not know the exact positions, or the arrival of englacial channels at the bottom, but glacier surface does not explain completely bedrock topography. Therefore, ice thickness measurements with GPR (Ground Penetrating Radar), done by Binder et al., 2006 were embraced in a semi-quantitative computation of hypothetical ice free terrain **Fig. 22**, chapter 6.1) By knowing the bedrock topography subglacial channels, as well as actual glacial outlets can be approximated. The travel time measurements in later experiments are easier explained by better approximation of the subglacial channel positions.

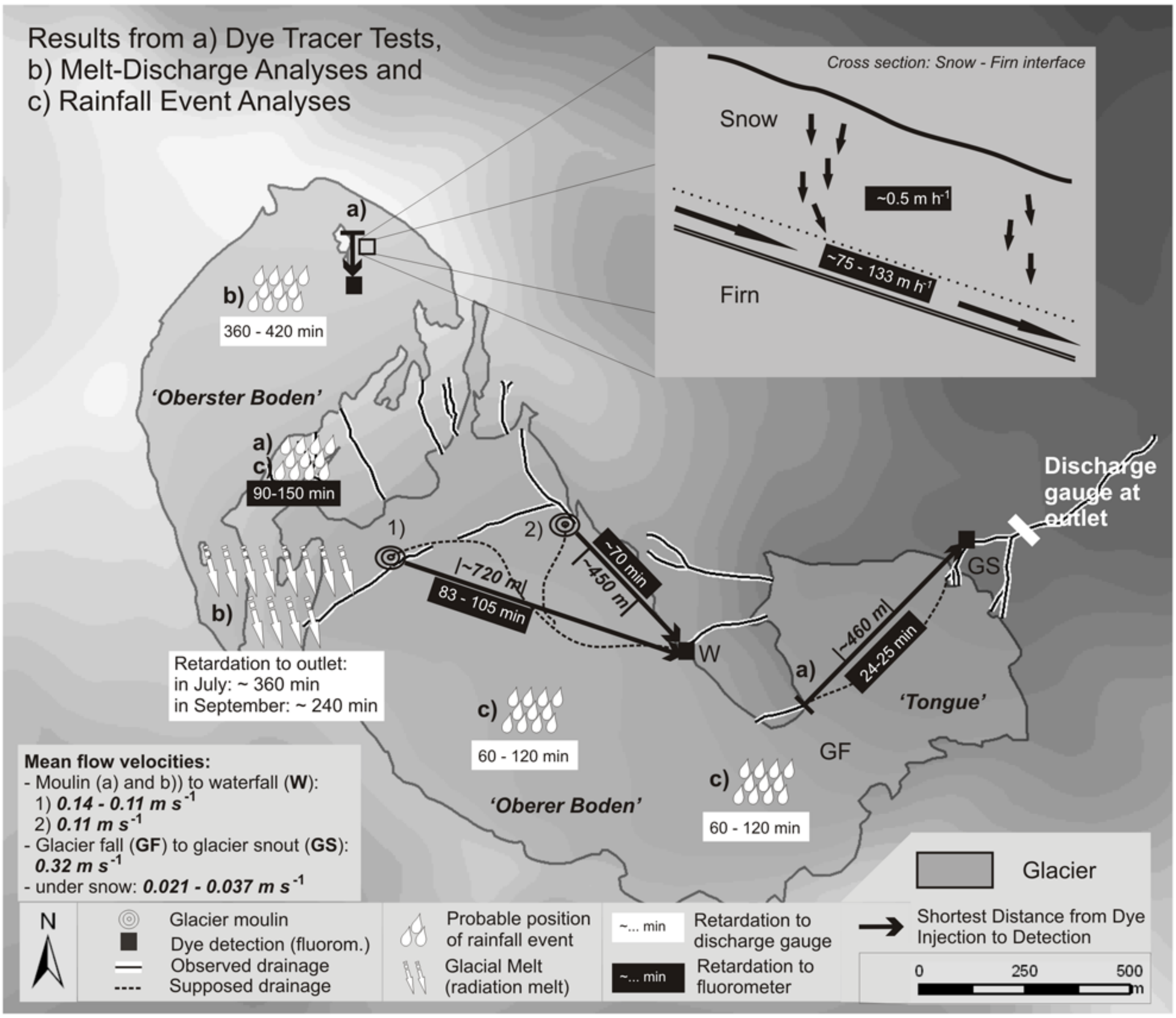


Fig. 58: Schematic view of the Goldbergkees showing a collection of gathered experimental and analytical results. *Source: own illustration*

Snowpack investigations with coloured dyes should illuminate processes of meltwater intrusion and water movement in the snow cover. In the beginning ablation season, Potassium Permanganate was sprayed on the still partly frozen snow pack. A series of profiles registered horizontal and vertical movement of water in snow. Melt and elevated temperatures induced snowmelt at the snow surface. Results showed that most of the water flows in a partly frozen snowpacks happens horizontally or parallel to the ground. The velocity with which water is mediated to the snow cover base depends on the point in the ablation period.

The beginning melt phase is characterized by frozen or at least partly frozen snow with denser snow layers. Those ‘textural horizons’ (Campbell et al., 2006), slow the vertical intrusion of meltwater expect for fractures. The further propagation of the melting front decomposes the layered texture of the snow. Whereas meltwater initially moves rather horizontally, it then percolates “stair- like” and finally vertically through the snow cover. (Fig. 28) The initially only locally restricted percolation of water into the snow is then observable over the whole plane. The decomposing metamorphosis of snow crystals is accompanied by the progression of the ablation season and results in an increase of percolation velocities.

For profiles at *Oberster Boden* horizontal flow rates between 0.08 m h^{-1} and 0.21 m h^{-1} (chapter 6.2.1.1) or 0.65 m h^{-1} (chapter 6.2.1.2) were measured.

Direct injections of coloured dye to draw water movement in deeper snow layers did not succeed as injection produced a remaining shaft functioning as an entrance for meltwater. The additional high water input rinses the injected dye (chapter 6.2.2).

In summary, water movement in snow is considerably slower than over firn or through the ice body showing good retention and storage characteristics. Nevertheless, the slow drainage through snow is accelerated during the ablation season. The increase of snow run-off velocities over the ablation season has not been quantified in this study.

Water movement through snow is of greater importance in the first third of the ablation season. The focus is relocated toward firn and glacial melt in the high ablation season. Therefore, fluorometric dye experiments investigating water movement on firn under snow and through the glacier body itself were undertaken.

Excavations of firn under snow showed the movement of water in the lowermost snow layers on firn. As long as no direct ruptures or crevasses are present, firn is hardly permeable. Meltwater produced at the snow surface flows plane or channeled at the top of it. Nevertheless it is not comparable with an open flow since the lowermost saturated snow layer builds an aquiferous stratum.

The injection of dye into this aquiferous layer and its detection registered ranges of meltwater movement velocities.

The measured velocities were between 75.58 and 133.3 m h^{-1} (0.0208 – 0.0369 m s^{-1}). Observations and comparison with other investigations showed highest flow velocities coinciding with higher flow rates *(compare Nienow, Sharp, Willis, 1996)*.

Flow velocities through the overlying snow could be deduced exemplarily by a single rainstorm event. The in-wash of particles from the snow surface could be registered as an elevated turbidity signal. A lag time of about 6 hours was calculated for a total depth of 3 m of snow for this time of the year. The results correspond to values of about 0.4-0.5 m h^{-1} given by *Campbell et al., 2006* for similar experiments. This leads to the conclusion that water is moving faster at a factor 10^2 - $2*10^2$ once it reached the firn table.

The main focus of this work though lies on the investigation of en- and subglacial channels (chap. 6.4). With the help of different fluorometric dyes, water movement is traced by injecting dyes into glacial moulins and into subglacial torrents. Two measurement campaigns in the high ablation season and at the very end of it, revealed aspects of water flow in en- and subglacial channels.

The englacial system was accessible by moulins and crevasses. Latter showed no superficial flow and were not adequate for a dye tracer input. On *Oberer Boden* mainly two moulins were observable. The amounts of flowing water as also courses of subglacial channels could only roughly be estimated. In fact, this caused problems in choosing the adequate tracer amounts. No indication of probable dilution of injected dyes could be done in the prefield. Moreover, elevated turbidity caused detection limits to rise and to influence dye tracer signals.

Travel times from the glacier moulin (see **Fig. 58**, moulin num. 2) to the detection device were around 70 minutes in high ablation season, corresponding to a mean flow velocity of 0.1086 m s^{-1}. Results for late ablation season measurements in field campaign 2 showed similar velocities. For a distance of 720 m from the moulin to the fluorometer a lag-time of 84 and 105 minutes was registered, corresponding to 0.1445 to 0.1122 m s^{-1}. *Nienow et al. (1996)* recorded velocities in the upper drainage system for Haut Glacier d'Arolla around 0.15 m s^{-1}. Moreover we can assume a subglacial branching of the channels as two different breakthrough curves were recorded for a single injection. Dye measurements in the tongue area showed faster flow velocities (0.3194 m s^{-1}). By turbidity measurements during rainfall, flow time between the uppermost areas and waterfall could be estimated, resulting in 90-150 minutes of lag-time. The mean flow velocities for the performed tracer tests are calculated according to shortest distances. We can expect though, actual en- and subglacial flow paths to be significantly longer. Exact positions of englacial and subglacial channels can only be estimated but could be approximated by ice thickness calculations (chap. 6.1).

On the basis of selected rainfall events in the summery ablation season, retardation effects were analysed. The selection was done for days without radiation melt, supported by data from precipitation stations and personal observations. Only few events could be clearly located and were strong enough to be traceable clearly at the discharge gauge. Nevertheless some events could be located with high probability to lower or upper parts of

the *Goldbergkees.* The hourly precipitation data resolution did only allow an analyze on an hourly base.
Lag-times for *Oberer Boden* and *Tongue* were between 1-2 hours, depending on the roughly estimated deposition location.
Rainfall events on *Oberster Boden* could be registered 6-7 hours after deposition at the discharge gauge.
The results from rain analyses should be classed with retardation times from melt analyses.

General system retardation was calculated by analyzing lag-times from dry melt days. The reaction of the discharge hydrograph mirrors the overall reaction to melt. Highest melt is assigned to highest discharge volumes. August retardation could not be determined since no complete dry day or proper melt phase was recorded. Moreover, melt and discharge volumes in October were too low for a proper evaluation.
Retardation in July was 6-7 hours whereas September lag-time dropped to 4 hours. Results represent an evidence for the transformation of glacial drainage systems during summer. Decreasing lag-times could not only be assigned to an ablation and the loss of snow but also to an increase in effectiveness of the drainage system. Short-term drainage of free melt or rainwater is reduced by the reduction of snow volumes. Entrances and moulins, as well as en- and subglacial passages are widened.

To sum up, gathered results about water storage and flow conditions on Goldbergkees are of good consistency.
Coloured dye tracing in snow at the beginning of the ablation period showed slow movement of water through partly frozen layers, the propagation of the melting front and the development of the in-snow percolation and drainage of meltwater through flow-fingers and between pores. Results reflected good temporal storage capabilities of barely structural decomposed snow.

Dye tracer experiments under snow in the high ablation season showed an increase of the percolation velocities through the snowpack, but show still the highest retardation in comparison with other glacial parts. The flow in the basal saturated snow layer is significantly faster and dependent on flow volumes. The collected meltwater from snow is transported to the entrances of the en- and subglacial drainage system.

Fluorometric dye tracer tests of en- and subglacial drainage reflected mean flow velocities of 0.1086 to 0.1445 m s^{-1} for Oberer Boden and 0.3194 m s^{-1} for discharge under the tongue. Moreover, indications of subglacial branching were found.

Discharge retardation analysis from rainfall events and dry melt days point to an increase of discharge velocities over the ablation period. In other words retardation characteristics shift to shorter term storage with the progression of the summer. The results from en- and subglacial drainage in special parts of the glacier fit into the general retardation values.

The results reflecting outputs of different investigation methods account for a general overview of the situation rather than a specific dimensioning of storage capabilities and flow behavior.

8 Perspectives

Results from the current study provided useful information about water storage and flow processes on the *Goldbergkees* glacier. Dye tracer tests investigating water movement on firn and in en- and subglacial conduits are of great value.

The general and *"broad"* approach including various investigation methods for a delineation of the complete ablation season ended in interesting results. Notwithstanding scientific approach, means and methods, own failures must be criticized in this context.

First of all, the approach resulted in great expenditure regarding the relation between effort and effective output. Different methods include great quantities of background theory, all to be reviewed in the run-up of the field experiments. Moreover, a complete investigation of different parts of the glacier drainage system was not always targeting. The particularities of the *Goldbergkees* basin impede a uniform handling of the basin.

Due to high risks, the glacier drainage system could not be sampled for some parts. The subdivision into two different basins and the break of ice cover between *Oberster Boden* and *Oberer Boden* complicated the field investigation as well. These subdivisions created unexpected problems and should be considered more carefully in following studies.

The application of colored and fluorometric dyes in glacial hydrology was appropriate. Thus, there are some problems to be considered:

- Application of sufficient dye tracer amounts
- Turbidity can influence the dye measurement significantly. Due to solved sediments and in-washes from the glacier surface, dye signals can be highly interfered.
- Dye tracing is only a punctual measurement regarding location and time.

The fact, that dye tracer experiments, if not done as a continuous injection, can be very problematic in a continuously changing environment. Hydrologic conditions vary with position, daytime, the annual cycle and meteorologic forces. Only by collecting continuous data over the complete ablation season at varying daytimes, a representative data base would have been established sampling over the complete ablation season. The expenditure in effort and means would have been too high.

The momentary record by dye tracing was therefore connected and completed by discharge analyses. But the representativeness of results was narrowed by the fact that outputs could not be easily compared. The changing drainage system expresses huge variations in run-off behavior.

Problems regarding the discharge analysis were caused by the high variability of rainfall. Controversial registration was made in some cases. Notwithstanding, the investigation area is one of the meteorologically best monitored high mountainous regions. This is why I would recommend any further investigations and studies of any hydrometeorologic or climatic background.

The good quality of hydro-meteorologic monitoring data for *Goldbergkees* region makes forget some problems related to accessibility and glacier shape characteristics.

I would not entirely recommend the selected methods for further investigations. Surely fluorometric dye tracing is an extensively approved mean in glacial hydrology. There is nothing against further dye tracer investigations, as long as expenditures in time and means are respected. The alteration of the drainage system over the summer forces a continuous sampling.

In fact, with greater time expenditure and greater financial and work-related means, it would be possible to measure storage quantities and time more meaningfully. A better preparation in the run-up would diminish mistakes in during measurement and applied dye tracer amounts.

Moreover, data output could be more precise if measurement would emphasize on selected aspects of storage capacities, weather that might be snow, firn, or ice. Surely, the components are linked and can not easily be observed without knowledge of active processes in one of the others components. Thus, a complete acquisition embracing snow, firn and ice is very difficult for the complexity of the drainage and storage system.

Finally, it is to say that despite all obstacles, further investigations in storage and drainage characteristics of glaciers could be of great value.

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