Pathways and transit time of meltwater in the englacial drainage system of Rabots Glacier, Kebnekaise, Sweden

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Preface

This Master’s thesis is Caroline Coch’s degree project in Physical Geography and Quaternary Geology at the Department of Physical Geography and Quaternary Geology, Stockholm University. The Master’s thesis comprises 60 credits (two terms of full-time studies).

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ABSTRACT.

Following the crash of a Norwegian Hercules aircraft on Rabots glaciär in the Kebnekaise mountain range in 2012, a field campaign was initiated in order to assess the fate of the hydrocarbon pollution in the system. It is hypothesized that soluble components of the aircraft fuel will be transported within the glacial meltwater. This thesis focuses on constraining the likely transit time and dispersion of the meltwater as a proxy for potential pollution pathways. Therefore, the hydrologic configuration of Rabots glaciär was studied during the ablation season 2013 by means of dye tracing experiments and discharge monitoring in the proglacial stream.

The analyses of the dye return curves and stream monitoring suggest that Rabots glaciär exhibits a widely efficient drainage system towards the end of the ablation season, but with analyses revealing heterogeneity in the drainage system form. The seasonal evolution of efficiency was also assessed, showing an increase over time, although was hampered by early onset of melting before the field season began. There are different hydrological configurations on the north and south side of the glacier, possibly influenced by shading. The system on the north side is routing meltwater along the glacier bed over a long distance as indicated by the turbid outlet stream. Water routing on the southern side likely occurs through englacial channels. This configuration may be influenced by the thermal regime and distribution of cold surface layers.

It has further been revealed that both systems are likely to be disconnected from each other. Pollution that is transported with the meltwater down from the crash site on the southern side does not reach the drainage system on the northern side. Besides revealing potential pathways for soluble hydrocarbon pollutants, this case study contributes to the previously very limited knowledge of Rabots glacial hydrology, and our general understanding of polythermal glacier hydrology.

Keywords: dye tracing experiments, glacier hydrology, hydrocarbon pollution
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LIST OF ABBREVIATIONS

AWS          Automatic weather station
BS           Braided system
c(t)         Concentration of dye at a particular time
CIR          Color infrared
CV           Coefficient of variation
D            Dispersion coefficient
d            Dispersivity
DEM          Digital elevation model
E            Nash-Sutcliffe efficiency
ELA          Equilibrium line altitude
GDEM         Global digital elevation model
GLOF         Glacial lake outburst flood
ibid.        Latin short for ibidem, refers to the preceding reference used
IOA          Index of agreement
IPR          Ice penetrating radar
NDSI         Normalized-difference-snow-index
NOAA         National Oceanic and Atmospheric Administration
NTU          Nephelometric turbidity unit
PBIAS        Percent bias
Q             Discharge
R²           Coefficient of determination
RJA          Northern outlet stream of Rabots glaciär
RJB          Southern outlet stream of Rabots glaciär
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<tr>
<td>RMSE</td>
<td>Root mean square error</td>
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<td>RSR</td>
<td>RMSE-observations standard deviation ratio</td>
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<td>RWT</td>
<td>Rhodamine water traceable</td>
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<tr>
<td>Sd</td>
<td>Standard deviation</td>
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<td>SIR</td>
<td>Shortwave infrared</td>
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<tr>
<td>SR</td>
<td>Storage retardation</td>
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<tr>
<td>SSC</td>
<td>Suspended sediment concentration</td>
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<td>T</td>
<td>Temperature</td>
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<td>$t_m$</td>
<td>Residence time, i.e. duration between dye injection and peak concentration</td>
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<tr>
<td>v</td>
<td>Mean throughflow velocity</td>
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<tr>
<td>W</td>
<td>Dye recovery</td>
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<td>x</td>
<td>Distance between injection point and sampling location</td>
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1 INTRODUCTION

Glaciers are natural reservoirs that store and release water at different time scales. They have been studied for a long time due to their significance for fresh water supply in rivers, sudden catastrophic drainage events, hydroelectric power production, global sea level variations, sediment transport, landform formation or glacier dynamics (e.g. Jansson et al. 2003). This range of study areas requires comprehensive knowledge and glacier drainage processes.

The glacier drainage system comprising of supraglacial, englacial, subglacial and proglacial water routing is highly variable in time and space and influenced by glacier geometry, thermal regime and climate. Dye tracing experiments have been used in a number of studies to investigate the hydrological configuration of glaciers (e.g. Seaberg et al. 1988; Willis et al. 1990; Hock and Hooke 1993; Fountain 1993; Nienow et al. 1998; Bingham et al. 2005). Using this technique, a tracer such as fluorescent dye or salt is injected into the drainage system and its concentration is subsequently recorded in the proglacial stream on emergence. The concentration-time plot (or dye return curve) contains information regarding the nature of the flow pathways and various calculations reveal properties such as transit time or tracer dispersion (Nienow and Hubbard 2005).

Following the tragic aircraft crash of a Norwegian Super Hercules aircraft into the west-facing wall of Kebnekaise, northern Sweden in March 2012, a monitoring program was initiated in order to assess the fate of the hydrocarbon pollution in the glacial system. The wreck and a considerable amount of aircraft fuel were buried in a snow avalanche triggered by the crash. Very few studies deal with hydrocarbon pollution in glacial environments, although oil and fuel spills are recognized as a common source of contaminants close to research stations in Antarctica (Bargali 2008). Most studies dealing with hydrocarbon pollution in polar environments have focused on the impacts on marine ecosystems and effects on soil (Colwell et al. 1977; Jordan et al. 1978; Song and Bartha 1990; Kennicutt II et al. 1991, 1992; Simpson et al. 1995; Aislabie et al. 1999; Cripps and Priddle 1991; Green et al. 1992; Jarsjö et al. 1994, 1997; Cripps and Shears 1997). Relatively little research has been done on the movement and fate of spilled fuel on ice (Tumeo and Larsson 1994; Jepsen et al. 2006; Brandvik and Faksness 2009; Faksness and Brandvik 2008).
The Rabots pollution monitoring program during 2012 and 2013 involved snow probing in order to understand volatilization and degradation processes and to assess the redistribution from the crash site. Water samples were collected in the proglacial stream of Rabots glaciär but also downstream of Björlings glaciär and Storglaciären that may be contaminated, too. As a proxy for the transport of pollutants, glacial hydrological methods were employed on Rabots glaciär to identify preferential flow pathways of melt water. While Rabots glaciär has been subject to previous studies focusing on radio-echo-sounding investigations to reveal the subglacial topography (Björnsson 1981) and on mass balance variations in response to climate (Stroeven and Van de Wal 1990; Brugger et al. 2005; Brugger 2007), the glacier’s hydrological configuration has remained poorly understood and little studied before now.

The objective of this thesis is thus to investigate Rabots glaciär’s hydrological configuration through undertaking and analyzing dye tracing experiments on the glacier and in the proglacial stream, as well as analyses for proglacial stream monitoring. In addition, the application of digital elevation models (DEMs) and Landsat 8 imagery allowed for the delineation of the glacier extent and the catchment as well as the derivation of hillshading over the course of a season. The study aims thus to understand both the catchment dynamics and the temporal development of the drainage system, producing results, which can help to constrain preferential transport pathways of the pollutants from the source zone. Moreover, study of Rabots glacier will contribute to a broader understanding of polythermal glacier hydrology and to verify current conceptual models.
2 THEORETICAL BACKGROUND

2.1 Thermal Regimes of Glaciers

According to the early geophysical classification by Ahlmann (1935), there are two main types of glaciers. (1) *Temperate* glaciers are at their melting point with exception of a cold surface layer up to a couple of meters that forms during winter. (2) *Polar* glaciers exhibit negative temperatures throughout the whole year and are further subdivided into *High-Polar* and *Sub-Polar* glaciers (ibid.).

This broad definition has been developed further as thermal configurations of glaciers were studied in more detail (e.g. Blatter and Hutter 1991; Pettersson 2004). At present, the common definition distinguishes three categories which are temperate, cold and polythermal glaciers. Temperate glaciers are found in maritime areas with high precipitation and summer melting, such as southern Iceland, western Norway or New Zealand. Cold glaciers are common in arid environments with small snow accumulation rates, for example Antarctica, whereas polythermal glaciers are widely spread over the earth (Benn and Evans 2010).

Polythermal structures of varying type have been observed in different locations. They are discussed by Blatter and Hutter (1991) and Pettersson (2004) and presented in Fig. 2-1. Types (a) and (b) are characterized by the formation of cold ice due to the cold conditions. Surface melt rates are typically very low, while warm ice can occur at the base due to strain heating. Examples for this thermal configuration are found in the Canadian Arctic (e.g. Copland and Sharp 2001), the Yukon Territory, Canada (e.g. Clarke et al. 1984) or Alaska, USA (Rabus and Echelmeyer 1997). Type (c) is typical for glaciers with a large elevation interval, where the accumulation area reaches down into warmer zones. Extensive melting lets meltwater percolate down through the firn pack, where it refreezes. This process leads to latent heat release, which warms the firm pack to temperate conditions. Hence, cold and temperate ice may both extend into the ablation area. The formation processes are similar for type (d), where cold ice occurs only in the upper part of the accumulation zone. Glaciers in the European Alps are often typified by this configuration (e.g. Haeberli 1976; Suter and Hoelzle 2002). Configuration (e) is characterized by predominantly temperate ice, except for a cold surface layer in the ablation zone. During the spring, the accumulation area becomes temperate when meltwater refreezes. The cold winters cause a layer of cold ice close to the surface in the ablation area, which is almost impermeable to water. Heat
penetration into the ground is limited, because ice is a poor conductor. If melt rates are small, this cold surface layer will remain cold throughout the year. Characteristic glaciers occur under the polar maritime conditions of Svalbard (e.g. Dowdeswell et al. 1984) or under more continental conditions on the eastern part of the Scandinavian mountains. Storglaciären is a well-studied example from Sweden (Pettersson et al. 2003, 2004; Pettersson 2004). Based on configuration (e), increasing net-ablation rates at lower elevations may remove major parts of the cold surface layer, while remaining in place at higher altitudes of the ablation zone (type f).

It is important to note that this categorisation provides only a description of the scope of possible polythermal configurations and does not cover all types that may be found. Polythermal glaciers may be predominantly cold (such as types a-c) or predominantly warm (types d-f) and show a range of individual configurations. As outlined before, just a few studies have focussed on the detailed thermal structure of polythermal glaciers, making it difficult to subscribe to one general conceptual model.

![Fig. 2-1. Range of thermal structures of polythermal glaciers from predominantly cold (a-c) to predominantly warm (d-f). The gray color displays temperate ice; ELA stands for equilibrium line altitude. See text for detailed descriptions (Pettersson 2004, based on Blatter and Hutter 1991)]
2.2 Hydrology of Polythermal Glaciers

A distinction is made between supra-, en-, sub- and proglacial water flow. These hydrological systems interact and are highly variable in time and space (Benn and Evans 2010). Figure 2-2 conceptually illustrates the glacial drainage system, showing supraglacial lakes and streams, swamp zones, moulins delivering water into the englacial and subglacial system, water filled crevasses, water filled fractures, subglacial channels and proglacial runoff.

The thermal structure of a glacier (described in chapter 2.1) can influence its hydrology. Glaciers with prevailing fractions of cold ice tend to develop mainly supraglacial and ice-marginal channels, because penetration of meltwater is by freezing. Warm ice allows meltwater to penetrate without considerable freezing and form supraglacial, englacial and subglacial channels. Thus, polythermal glaciers develop very complex hydrological systems (Bennett and Glasser 2009). Connections between supraglacial and subglacial drainage systems on glaciers with cold ice have been thought to be limited for some time. However, a study by Bingham et al. (2006)

![Fig. 2-2. Schematic view of the glacier drainage system comprising of supraglacial features, such as surface lakes (A), supraglacial streams (B) and swamp or slush zones (C); englacial features, such as moulins leading to the subglacial system (D), water filled crevasses transporting water within the glacier (E) and water filled fractures (F); subglacial features (G) and proglacial runoff (H) (modified from Cuffey and Paterson 2010).]
revealed indications for a short-term and/or seasonal evolution of an englacial and subglacial drainage system on the predominantly cold polythermal John Evans Glacier, High Arctic Canada.

This section is going to follow the path of meltwater from source and storage mechanisms through water transport on, through, under and in front of the glacier. Case studies are mainly taken from polythermal glaciers.

2.2.1 Water Supply and Storage

There are four sources for liquid water in glaciers according to Paterson (1971) and Lliboutry (1976): (1) melting from strain heating; (2) adaptation of the pressure-melting point based on modifications in lithostatic pressure; (3) surface water from melt processes entering the glacier via cracks and crevasses in the ablation zone and (4) water caught in the ice at the transition between firn and ice in the accumulation zone. The first two sources can be quantified by modeling, but the latter two sources are hard to estimate because of the complex processes (Aschwanden and Blatter 2005).

Jansson et al. (2003) review storage mechanisms of glaciers, that appear as water, snow and ice. They are particularly important when evaluating water supply on different time scales. Long-term storage applies to storing ice and firn on time scales of years to centuries. Intermediate-term storage takes place on a seasonal scale and includes storage and release of snow and water. Diurnal drainage processes through the glacier proceed on the time scale of short-term storage. Moreover, sudden events such as surges or glacial lake outburst floods (GLOF) that occur irregularly, are put in the category of singular storage releases. This integrated approach emphasizes the variations of water supply and helps to model runoff of a glacierized catchment.

2.2.2 Supraglacial Water Flow

Water flow on a glacier’s surface varies due to the different permeabilities of ice and snow (Benn and Evans 2010). At the end of the ablation period, snow and firn are exposed to the surface in the accumulation area at higher elevations whereas bare ice is existent in the ablation zone. The equilibrium line marks the boundary between accumulation and ablation zones (Fountain and Walder 1998). The processes of snow melt and water routing are described below.
Firn plays a significant role in temporary water storage as various studies emphasize. Fountain (1989) studied the storage and hydraulic characteristics of firn on South Cascade Glacier, Washington State, USA and found that the firn stores a significant amount of water from the surface and controls its drainage into the body of the glacier via crevasses. Fountain and Walder (1998) further propose that glacial drainage is delayed or ‘buffered’ by storage in firn, and that drainage through snow and firn in the accumulation area is the source of base flow of glacial runoff. The development of a firn aquifer had been recognized in numerous studies, including investigations of water movement in firn on Storglaciären, Sweden by Schneider (1999) and the assessment of the significance of the firn layer for glacial discharge on Hofsjökull, Iceland by Woul et al. (2006).

Liquid water, derived from precipitation or energy exchange between atmosphere and glacier, percolates through the snow pack and will refreeze. The release of latent heat causes warming of the snow cover up to the melting temperature. This snow is also described as wet snow. A study on Haut Glacier d’Arolla, Valais, Switzerland by Campbell et al. (2006) emphasized the importance of the supraglacial snowpack in mediating melt water supply to the glacier system. The snowpack acts as a short term storage component, showing complex flow patterns and ice layers that delay percolation. An increase in drainage efficiency through the snowpack over the season had been observed.

Hodgkins (2001) also observed water retention at the beginning of the ablation season at the non-temperate Scott Turnerbreen glacier on Svalbard, where its relatively flat surface results in a low hydraulic gradient and allows the formation of slush zones with puddles of standing water. The release of supraglacially stored meltwater therefore dominated the hydrograph and obscured the diurnal forcing.

Drainage occurs laterally through interconnected pores in the snow, creating rills and finally incising channels on the surface with rising water content (Benn and Evans 2010). These processes become more efficient over the season (Jansson et al. 2003). An example of a supraglacial stream, taken from Rabots glaciär, Sweden, is presented in Fig. 2-3. The rate of downcutting of a supraglacial stream into the ice surface can be calculated based on a model for open channel conditions by Fountain and Walder (1998), where incision rates are directly related to channel slope but less dependent on discharge and channel roughness. These theoretical
assumptions are confirmed by field observations by Gulley et al. (2009a), who investigated the behavior of supraglacial streams at the pycnothermal glaciers Longyearbreen, Svalbard and Khumbu Glacier, Nepal. They further recognized extensive supraglacial drainage networks only in uncrevassed regions of the glaciers. Benn and Evans (2010) state that the roughness of the ice wall channels influence flow velocities within the stream, such that low roughness will result in high flow speeds. The water will either drain directly at the ice margins (e.g. Hodgkins et al. 2009) or flow into the englacial system as described in the following section.

2.2.3 Englacial Water Flow

The hydrology of various glaciers from Svalbard, concerning their thermal regime and hydrological characteristics was reviewed by Hodgkins (1997). He revealed that englacial water routing does occur although thought to be absent on non-temperate glaciers for a long time. Numerous subsequent studies deal with the connection between supraglacial and subglacial hydrology (Boon and Sharp 2003; Copland et al. 2003; Bingham et al. 2006; Das et al. 2008; Gulley et al. 2009; Catania et al. 2010). This section is going to examine the current state of knowledge concerning englacial conduit formation.

Shreve (1972) made a first attempt in modeling englacial drainage, basing the model on three principal assumptions: (1) steady state of the englacial system, (2) normal flow to equipotential surfaces and (3) the formation of arborescent passage networks with time. However, this model
could not fully describe the development of drainage networks as discussed by Gulley et al. (2009b). For example, steady state conditions are practically absent, because surface melt varies diurnally and annually. Further, the assumption that recharge is evenly distributed across the glacier’s surface is not met since discharge is accumulated at just a few points. This causes other equipotential surfaces than predicted in the Shreve model. There are occasions where the model predicted subglacial conduits at correct locations, for example on Haut Glacier D’Arolla by Sharp et al. (1993), but most likely for the wrong reasons as Gulley et al. (2009b) conclude. The authors review various mechanisms of englacial conduit development, which are backed by direct observations (1) cut-and-closure of supraglacial streams, (2) routing through permeable structures and (3) hydrofracturing.

One conceptual model for englacial channel formation, referred to as the cut-and-closure mechanism, has been studied on Longyearbreen (Gulley et al. 2009a) and Austre Bøggerbreen (Vatne 2001), Svalbard. Supraglacial channels incise into the ice surface which creep closes the channel roof. This is assumed to be the prevalent mechanism for channel formation in uncrevassed areas of polythermal glaciers. Conduit formation by cut-and-closure needs high stream incision rates (favoured in regions of high discharge and surface slope) and low ablation rates (found in cold environments or on debris covered glaciers). The connection between the supraglacial and subglacial systems is regarded as relatively inefficient (ibid.).

Permeable structures in glacier ice that can be exploited may be debris filled crevasses or fractures. Gulley and Benn (2007) investigated three valley glaciers in the Khumbu Himal, Nepal, that exhibit extensive debris covers. It turned out that relict debris filled crevasse traces feature a high hydraulic conductivity and provide the basis for conduit development. The conduits grow by melting of the ice walls and draining of the debris (ibid.). Fountain et al. (2005) explored the hydrological system of Storglaciären, Sweden through video images from high pressure hot borehole drills. At all depths, including close to the glacier base, they found hydraulically connected fractures. Concerning the spatial extent, this fracture network is dominant of englacial drainage (ibid.). However, according to dye tracing investigations by Seaberg et al. (1988), most point-sourced supraglacial meltwater reaches the glacier bed via moulins. Gulley (2009) mapped englacial conduits on the temperate Matanuska Glacier, Alaska, USA. He distinguished between two broad conduit categories based on the orientation of the principal stresses: (1) conduits developed at locations where a big supraglacial stream met longitudinal crevasses leading to
hydrostatic crevasse propagation; and (2) conduits formed due to the sinking of smaller streams on shear crevasses (ibid.).

Hydrologically driven fracture propagation as a very efficient mechanism has been described by various authors (e.g. Boon and Sharp 2003; Alley et al. 2005, van der Veen 2007). Under conditions of sufficient tensile stresses and water supply (Weertman 1973), fracture propagation can potentially reach the glacier base. Lakes form in topographic depressions on the ice surface and are usually recharged by supraglacial streams. Overdeepening of longitudinal crevasses by supraglacial streams or lakes has also been observed to force compressive hydrofracturing that may propagate down to the bed (Gulley et al. 2009b). Research on the predominantly cold John Evans Glacier, Ellesmere Island, Arctic Canada has shown that meltwater from the surface propagates through the glacier and reaches the bed, affecting horizontal and vertical surface velocities significantly (Copland et al. 2003). This rapid increase in surface velocities support the theory of hydrofracturing as outlined by Boon and Sharp (2003) and Zwally et al. (2002).

2.2.4 Subglacial Water Flow

Very little is known about the specific subglacial environment of High Arctic glaciers (Irvine-Fynn et al. 2011). However, theories about subglacial hydrology of temperate glaciers exist and may be applicable to polythermal environments. These are now discussed in the following paragraphs. It is important to note that the theories have limited observational evidence.

Basal melting or limited meltwater supply from a glacier surface may form a thin water layer at the bed of a glacier, termed a Weertman film (Weertman 1972). The water layer develops as a result of locally increased melt rates, for example on the upglacier side of bumps on the bed. In turn, the down-glacier side of a bump experiences refreezing, also known as regelation. A water film may act as a lubricant, acting to promote basal sliding (Bennett and Glasser 2009).

Efficient transport through channels that cut upwards into the ice are referred to as Röthlisberger or R-channels (Röthlisberger 1972). The frictional energy due to water routing causes enlargement of passages and prohibits closing by inward ice flow. The balance of both processes, enlarging and closing, influences the tunnel size, and the theory is based on the assumption of steady flow. The drainage network presumably covers a limited area of the glacier bed in such dendritic networks (Fountain and Walder 1998; Walder 2010).
Subglacial cavities are thought to develop when glacier ice slides across a rough surface, causing the ice to separate from the glacier bed. Depending on the bedrock topography these cavities may cover extensive areas of the glacier bed. They may be connected by a network of small links, also termed Nye or N-channels, usually cutting down into the bedrock or sediments or by small R-channels. Generally, this drainage configuration is assumed to be relatively inefficient, showing distributed flow at low transit times (Bennett and Glasser 2009).

Glacial till underneath a glacier can act as subglacial aquifer. Water may move through pore spaces along the hydraulic potential gradient (Darcian flow), through small channels and pipes or as a thin water layer on top of the sediment surface (Bennett and Glasser 2009).

Distributed drainage systems where meltwater is widespread over the glacier bed are distinguished from discrete drainage systems where meltwater is routed through tunnels and channels across the bed. N-channels as well as R-channel networks are categorized as discrete drainage systems. The Weertman film is assumed to be a distributed system whereas flow through subglacial sediments can occur on both configurations (distributed as Darcian flow or discrete through tunnels and pipes). Linked cavity systems are also prone to show distributed flow (Benn and Evans 2010). In general, channelized systems are more efficient pathways for evacuation of water from a glacier than distributed systems. Furthermore, subglacial configuration is subject to spatial and temporal changes as the ablation season progresses and meltwater inputs increase (Bennett and Glasser 2009). For example, a study by Cowton et al. (2013) at Leverett Glacier, a land-terminating outlet glacier of Greenland, documented the rapid transition from distributed to channelized drainage by means of dye tracing experiments.

2.2.5 Proglacial Water Flow

Glacierized catchments are characterized by the existence of snow and ice and the common lack of soil and vegetation that results in high surface runoff. (Willis 2005). As Röthlisberger and Lang (1987) outline, the glacier runoff regime is either controlled by meltwater, precipitation or both. Catchments with a glacier cover of about 40% show the least variability in runoff. This is due to the fact that precipitation generally reduce meltwater runoff because of less incoming solar radiation. The authors refer to this process as the compensating effect. (ibid). As seen in Fig. 2-4, runoff variability then increases as the percentage of glacier cover both rises and falls (Willis 2005).
Willis (2005) further examines how annual runoff from glacierized catchments is linked to different meteorological variables, such as intensity of radiation, air temperature, precipitation, humidity and cloudiness. There are therefore significant variations depending on the geographical setting and local climate. The following paragraph summarizes the specific runoff characteristics for different climatic regimes.

Generally, midlatitude glacierized basins (example Fig. 2-5a) show long summer melt seasons that are connected to mean daily temperatures exceeding 0°C. The large annual runoff variation occurs because rainfall events coincide with the summer melt. Furthermore, temporary water storage in the beginning of the melt season causes a delay in timing of the annual peak flow. Annual runoff at high latitudes is largely dependent on fluctuations in solar radiation and air temperatures. The melt season is commonly shorter and fluctuations are typically lower compared to the midlatitudes. Figure 2-5b shows an example for Svalbard with a maritime influence while Fig. 2-5c presents runoff from a glacier in Antarctica. Glacierized basins at low latitudes are exposed to much lower fluctuations in solar radiation and air temperature; however, they experience larger variations in humidity, cloudiness and precipitation. Typically they are found at high altitudes. Distinctions are made between (1) summer accumulation types with dry, cold winters and pronounced summer precipitation and (2) year round ablation types with high temperature, radiation and precipitation rates throughout the year. Figure 2-5d shows an example from Bolivia, where the highest proportion of annual runoff appears between December and March (summer melt). Runoff during the winter months of June and July in this case is related to precipitation (Willis 2005).
According to Röthlisberger and Lang (1987) base flow consists of (1) groundwater flow, (2) storage released water (snow, firn, cavities, lakes) as well as (3) subglacial meltwater. These components are relatively constant on a daily basis. Water derived from melting or precipitation is superimposed on the base flow and composes the diurnal cycles (ibid.).

Maximum discharge generally occurs a few hours after the maximum melt rate, whereas the time lag decreases as the ablation season proceeds. The dynamic changes are presented in the case study of Vernagtbach in the Austrian Alps (Fig. 2-6). The drainage area covers 11.44km\(^2\) with a glacier coverage of 84%. It is clear that both, discharge amount and amplitude increase with time and that peakflow tends to occur earlier in the day as the season progresses. This reflects the increasing efficiency of the drainage system (Röthlisberger and Lang 1987; Swift et al. 2005, Benn and Evans 2010; Cuffey and Paterson 2010). High melt rates occurring in the beginning of the ablation season result in high meltwater supplies and may lead to high velocity or spring events. These events have been observed on Haut Glacier d’Arolla, Switzerland (e.g. Mair et al. 2003). The high bulk discharge and high subglacial water pressures cause an ice-bed decoupling and are important for glacier dynamics (ibid.).

Fig. 2-5. Runoff from various glacierized catchments representing different intraannual variations depending on their geographic location (Willis 2005).
Annual, seasonal and diurnal discharge variations are not just limited to midlatitude glacierized catchments. The study of Mittivakkat Glacier in southeast Greenland by Mernild et al. (2006) recorded a time lag between maximum melting and discharge. Throughout the ablation season it decreased from between 5 and 7 hours in May to between 3 and 4 hours in August (ibid.). A study of the proglacial zone of Finsterwalderbreen, Svalbard by Hodgkins et al. (2009), another high latitude example, investigated the annual runoff cycle, which is closely connected to air temperatures and consequent melt rates. Also a review on water balance studies in Svalbard by Killingtveit et al. (2003) confirm the relation between melt rates and temperature. They showed that discharge in the early ablation season (June and July) is derived from snowmelt; later in the season (August and September) rainfall and surface melt dominate runoff of (ibid.).
2.3 The Dye Tracing Method

Dye tracing experiments provide an indirect method to investigate the drainage network of a glacier. A fluorescent dye is injected into flowing meltwater, either at the ice surface (via crevasses, supraglacial streams or moulins) or directly into englacial or subglacial system (via boreholes). A fluorometer records the concentration of the returning dye in the proglacial stream at the glacier snout. The obtained time-concentration curve (breakthrough curve) enables inferences regarding the characteristics of the glacial drainage system to be made (Hubbard and Glasser 2005; Leibundgut et al. 2009; Nienow 2011).

There are different types of fluorometers, which require careful handling. First, handheld fluorometers are small in size and may directly be used in the field. They may have a limited capacity of channels and since handling in the field may be difficult due to harsh weather conditions, the samples may be analyzed in the laboratory later on. Second, there are fluorometers that record the dye concentration automatically. They usually consist of a measuring sonde, which connects with a signal cable to a datalogger commonly in a waterproof box. For data acquisition, the sonde is immersed into the stream, while the logger records time and dye concentration automatically. Generally, these instruments allow for adjustment of the sampling intervals and number of channels recording. They can record several tracers simultaneously by measuring the wavelengths of the particular tracers (Schnegg 2003).

Fluorescence tracers are categorized as artificial, solute tracers. Their name originates from the fluorescence that is created when a substance absorbs light at one wavelength and emits it at a different wavelength. Tracers that are commonly used in glaciology are Rhodamine WT or RWT (water traceable) and Uranine. RWT has a pink/red color and is water soluble. The detection range is at wavelengths between 530 and 555 nm. Uranine is derived from the substance fluorescein, a dark orange powder, and transforms into a green-yellowish solution. The detection range is between 494 and 521 nm. Caution needs to be exercised because Uranine is extremely sensitive to sunlight and pH. It may therefore undergo fast photochemical decay, and is best suited to direct injection into the englacial system. Figure 2-7 shows the injection of Uranine 33.3% directly into a moulin. RWT and Uranine can be applied simultaneously because they cover different detection ranges (Leibundgut et al. 2009).
Dye tracing experiments have been applied in a range of glacial environments (e.g. Seaberg et al. 1988; Willis et al. 1990; Hock and Hooke 1993; Fountain 1993; Nienow et al. 1998; Bingham et al. 2005). Based on the dye return curves and calculations, characteristics of the glacier drainage system are revealed. The parameters transit distance, residence time, transit speed, dispersion, dispersivity and dye recovery are going to be examined in more detail in section 4.2.3.

Fig. 2-7. Injection of Uranine 33.3% into a moulin on Rabots glaciär, Kebnekaise, Sweden. Flowing water is crucial for a successful dye tracing experiment to avoid tracer loss by freezing onto the ice (Photograph by Caroline Coch).
3 RABOTS GLACIÄR

3.1 Setting

The valley glacier Rabots glaciär is located on the western part of the Kebnekaise massif, northern Sweden at N 646510 m E 7536919 m, SWEREF99 TM (Fig. 3-1). It consists of three accumulation areas, split into cirques oriented WSW, WNW, N and one ablation area. Its maximum recorded Holocene extent dates back to 1916 and has been continuously decreasing since then (Karlén 1973). The areal extent during summer 2013 was approximately 3.08 km², with a maximum elevation of 1848 m a.s.l. and a minimum elevation of 1111 m a.s.l. The average slope is calculated at 11.47° with a minimum of 1.22° in the ablation zone and a maximum of

![Fig. 3-1. Overview of Rabots glaciär. (a) Position within Scandinavia (Esri 2013). (b) Photograph from a helicopter (provided by Per Holmlund). (c) Map showing 2013 glacier extent (based on Landsat 8 satellite image), contour lines from 2011 (DEM provided by Keith Brugger) and outlet streams (based on GPS coordinates taken in the field in 2013). A detailed description of the methods employed is provided in sections 4.2.7 and 4.2.8.](image-url)
44.95° high up in the cirques. The glacier is presumably polythermal with the presence of a cold surface layer (Stroeven and Van De Wal 1990), much like the well-described 30-40 m thick cold surface layer of Storglaciären (e.g. Holmlund and Eriksson 1989).

There are two main proglacial streams, the northernmost of which is the most turbid, suggesting subglacial origin; while the southern stream is relatively clear. They both merge approximately 145 m away from the glacier front. Some smaller streams also exiting the glacier, however being much smaller in size and mixing with the main outlet streams. The proglacial environment is characterized by several braided systems routing along the local topography, which is characterized by two overrun moraines and the pronounced terminal moraine (Karlén 1973). The overall catchment size amounts 9.42 km² with an ice covered area of 32.67%.

Temperature and precipitation data presented in Fig. 3-2 show the prevailing conditions during the 2013 study period of this project. The second diagram from the ablation season of 2013 as it shows high temperatures at the end of May/beginning of June, which caused an early onset of

![Climate data retrieved from the automatic weather station (AWS) on Rabots glaciär showing precipitation and temperature for the ablation seasons 2012 (a) and 2013 (b). Note the high temperatures in early June 2013. (graphs provided by Pia Eriksson)
melt. The high volumes of melt water allow the presumption that the drainage system was already well developed by the start of the monitoring in July and August. Three particularly pronounced rainfall events could also be discerned from the 2013 ablation season data.

### 3.2 Research on Rabots glaciär

In 1883, the French geographer Charles Rabot documented Rabots glaciär for the first time during a period of presumed advance. Since then, a relatively few studies have been conducted on the glacier in comparison to nearby Storglaciären. Schytt (1959) presented a preliminary glacier inventory of the Kebnekaise massif including topographic characteristics and regime investigations. Karlén (1973) explored glacier and climate variations in the Kebnekaise Mountains during the Holocene in more detail, concluding that Rabots glaciär reached its Holocene maximum extent in 1916, documented in the landform record by the terminal moraine ridge. Mass balance measurements on the glacier are available since 1981 and provided by Tarfala Research station (Fig. 3-3).

Björnsson (1981) investigated the subglacial topography of Rabots glaciär by means of radio-echo-sounding, providing a longitudinal line along the center and nine traverse lines. The study found an area of 4.1 km², ice volume of 346 x 10⁶ m³, average ice depth of 84 m and a maximum ice depth of 175 m. In contrast to the other studied glaciers, Isfallsglaciären and Storglaciären, on the Kebnekaise massif, no basins (overdeepenings) were found under Rabots glaciär with this technique (ibid.).

![Fig. 3-3. Mass balance data for Rabots glaciär. The white bars represent winter balance while the grey bars represent summer balance. The net balance is marked in blue (positive) and red (negative) respectively. (Tarfala Data; Bolin Centre, http://bolin.su.se/data/tarfala/rabot.php, 6 Feb., 2014).](image-url)
The mass balance and flow regimes of Storglaciären and Rabots glaciär were studied and compared by Stroeven and Van De Wal (1990) with results suggesting that the net balance records of both glaciers were generally well correlated. However, while Storglaciären presumably reached its steady-state around the mid-1980’s, Rabots glaciär continued to retreat. It is assumed that this is caused by the geometry; Rabots glaciär is flatter at the ELA in comparison with Storglaciären. The authors proposed that also Rabots glaciär would continue to adjust to the former warming and reach equilibrium. Brugger et al. (2005) and Brugger (2007) continued in this research area by looking at glacier length variations and conducting glacier response modelling. Their results suggested that changes in length, volume and ice thickness on Rabots glaciär are significant and not in balance with the present climate. As mentioned in previous studies, Storglaciären had stabilized and found its equilibrium. This further supports non-synchronous behaviour of both glaciers in response to the 20th century climate (Brugger et al. 2005). The time-dependent numerical model applied by Brugger (2007) suggests differences in response time of both glaciers. Concerning volume changes due to small perturbations, the response time of would be ~215 years for Rabots glaciär and ~125 years for Storglaciären respectively. Response time in terms of glacier length changes was modelled to be 1.5 times longer for Rabots glaciär than for Storglaciären. It is important to note that the model gives an idea of the relative magnitudes of response time differences, rather than providing certain numbers of individual response times. The study associates these differences with the diverse longitudinal geometries, velocity profiles and the particular gradients of net balance rather than varying microclimates (Brugger 2007).

To date, there is no published data available concerning the detailed thermal configuration of Rabots glaciär. Field campaigns employing ground-based ice penetrating radar (IPR) and airborne radar had been conducted several times; however data processing remains ongoing. Nevertheless, there is already evidence for the presence of a cold near surface layer on Rabots glaciär (Peter Jansson, pers.comm., 4 Dec., 2014; Per Holmlund, pers. comm., 16 Oct., 2014). Another area of research that has lacked focus on Rabots glaciär is the glacial hydrological system. This study will add to the previously limited knowledge of the glaciers hydrological system.
4 METHODOLOGY

4.1 Field Methods

A field campaign was carried out during the ablation season of 2013 with a series of dye tracing experiments on the glacier and in the proglacial stream. These tracer experiments are considered as proxies for potential pollution pathways. An overview of all dye tracing experiments is presented in following table (Table 4-1) including the assigned code, date, location, type and amount of dye and whether a return curve was obtained. During some experiments the FL30 instrument also recorded turbidity data, which is also listed with the particular dates and times. Snow sampling was also conducted within the project to assess the behaviour of the hydrocarbon pollution in the accumulation zone. The reader is referred to the thesis by Ane LeBianca, where those results are discussed in detail.

The general procedure of performing dye tracing experiments is briefly summarized here. Before going to the field, calibration standards of each dye as well as bottles containing experiment

Table 4-1. Overview of dye tracing experiments on the glacier and in the proglacial stream. J stands for July glacier dye tracing experiments, A for August glacier dye tracing experiments and D for discharge measurements in the proglacial stream.

<table>
<thead>
<tr>
<th>Code</th>
<th>Date</th>
<th>Location</th>
<th>Dye</th>
<th>Amount [ml]</th>
<th>Sampling</th>
<th>Return curve</th>
<th>Simultaneous turbidity measurements</th>
</tr>
</thead>
<tbody>
<tr>
<td>J1</td>
<td>2013-07-04</td>
<td>Supraglacial</td>
<td>RWT 20%</td>
<td>125</td>
<td>auto</td>
<td>Yes</td>
<td>-</td>
</tr>
<tr>
<td>J2</td>
<td>2013-07-05</td>
<td>Crevasse</td>
<td>RWT 20%</td>
<td>245</td>
<td>auto</td>
<td>Yes</td>
<td>-</td>
</tr>
<tr>
<td>J3</td>
<td>2013-07-22</td>
<td>Moulin</td>
<td>RWT 20%</td>
<td>250</td>
<td>auto</td>
<td>Yes</td>
<td>-</td>
</tr>
<tr>
<td>A1</td>
<td>2013-08-08</td>
<td>Moulin</td>
<td>RWT 20%</td>
<td>150</td>
<td>auto</td>
<td>Yes</td>
<td>16:30 - 22:00</td>
</tr>
<tr>
<td>A2</td>
<td>2013-08-08</td>
<td>Moulin</td>
<td>Uranine 33.3%</td>
<td>100</td>
<td>auto</td>
<td>Yes</td>
<td></td>
</tr>
<tr>
<td>A3</td>
<td>2013-08-09</td>
<td>Moulin</td>
<td>RWT 20%</td>
<td>150</td>
<td>auto</td>
<td>Yes</td>
<td>9:30 - 21:00</td>
</tr>
<tr>
<td>A4</td>
<td>2013-08-09</td>
<td>Moulin</td>
<td>Uranine 33.3%</td>
<td>100</td>
<td>auto</td>
<td>Yes</td>
<td></td>
</tr>
<tr>
<td>A5</td>
<td>2013-08-10</td>
<td>Crevasse</td>
<td>Uranine 33.3%</td>
<td>200</td>
<td>auto</td>
<td>No</td>
<td>9:00 - 23:30</td>
</tr>
<tr>
<td>A6</td>
<td>2013-08-10</td>
<td>Supraglacial</td>
<td>RWT 20%</td>
<td>125</td>
<td>auto</td>
<td>No</td>
<td></td>
</tr>
<tr>
<td>A7</td>
<td>2013-08-15</td>
<td>Supraglacial</td>
<td>RWT 20%</td>
<td>500</td>
<td>auto, manual</td>
<td>Yes</td>
<td>11:30 - 23:30 00:00 - 18:00</td>
</tr>
<tr>
<td>D1</td>
<td>2013-08-08</td>
<td>River</td>
<td>RWT 20%</td>
<td>30</td>
<td>auto</td>
<td>Yes</td>
<td>see above</td>
</tr>
<tr>
<td>D2</td>
<td>2013-08-10</td>
<td>River</td>
<td>RWT 20%</td>
<td>30</td>
<td>auto</td>
<td>Yes</td>
<td>see above</td>
</tr>
<tr>
<td>D3</td>
<td>2013-08-15</td>
<td>River</td>
<td>RWT 20%</td>
<td>60</td>
<td>auto</td>
<td>Yes</td>
<td>see above</td>
</tr>
<tr>
<td>D4</td>
<td>2013-08-18</td>
<td>River</td>
<td>RWT 20%</td>
<td>50</td>
<td>auto</td>
<td>Yes</td>
<td>-</td>
</tr>
<tr>
<td>D5</td>
<td>2013-08-18</td>
<td>River</td>
<td>RWT 20%</td>
<td>100</td>
<td>auto</td>
<td>Yes</td>
<td>-</td>
</tr>
</tbody>
</table>
specific tracer injection volumes were prepared. In the field, the FL30 instrument was set up carefully at a location ideally 30 cm below the water surface to avoid scattering of daylight. To adjust the instrument to the stream temperature, the sonde was immersed in the stream for several minutes before conducting calibration. A calibration solution of 100 ppb was prepared in a 10l bucket. After performing the calibration and noting the time difference between logger and local time, the data acquisition could begin. The instrument allows an adjustment of the number of channels in use and sampling intervals. In the case of manual water sampling, the samples were collected and analysed in the laboratory at a later time with the handheld fluorometer when the samples had adjusted to room temperature. Further, the handheld instrument was recalibrated after every five samples to avoid a drift in the results. To mitigate the influence of scratches or dust on the cuvettes, they were cleaned carefully in advance. Additionally, every sample was measured four times in each orientation. Flowing water is crucial for the injection location to avoid freezing onto the ice.

4.1.1 Glacier Dye Tracing Experiments

The location of the FL30 automatic sampler differed in July and August as depicted in Fig. 4-1. In July, it was placed close to the glacier front where both main outlet streams merged whereas it

---

Fig. 4-1. Map of Rabots glaciär showing the glacier dye tracing experiments in July and August. Successful experiments are marked in black whereas experiments yielding no return are marked in grey. Both main outlet streams merge together close to the glacier front, where the FL30 experiment was placed during July. In August, the instrument was moved further downstream and placed by the gauging station.
was placed further downstream by the gauging station during August. The decision to move the instrument in August was made such that a longer travel distance would allow full mixing of the dye. Moreover, dye tracing experiments in the river could be performed at this location to produce a rating curve for calculation of discharge. The injection locations for the July experiments varied in the type of injection site used as no active moulins could be found in early July, with one injection into a supraglacial stream (J1), one into a crevasse (J2) and one into a moulin later in July (J3). Except for experiment A7, all August experiments were conducted at lower elevations of the glacier into active moulins. The simultaneous use of Uranine 33.3% and RWT 20% enabled the execution of two experiments per day: A1 and A2; A3 and A4 respectively). No return curve was obtained for the experiments A5 and A6. Reasons for that are going to be discussed at a later point. In addition to automatic recordings, manual sampling was performed in both outlet streams during experiment A7 in order to investigate the possibility of a preferential drainage pathway through the glacier system.

4.1.2 River Dye Tracing Experiments

Five dye tracing experiments were conducted in the proglacial stream to obtain discharge values. This was done on different days in order to capture various weather and runoff conditions, aiming to measure at both high and low stage levels. The amount of RWT 20% injected varied between the experiments due to changing hydraulic conditions and estimated discharge.

4.1.3 Field Instrumentation

The river gauging station comprised of a sonic radar sensor that measures the distance from the sensor to the target (river surface), and an instrument, which records meteorological variables (humidity and air temperature). The discharge station program runs every minute and stores an average value of each variable every 15 minutes. Detailed information on the instrumentation and logger program are provided in the manual by Peter Jansson. The instrument gives the relative gauging height, recording fluctuations in water level (stage). The gauging height monitoring was carried out between 2013-07-22 and 2013-09-06.

Another instrument recorded air and water pressure at a 10-minute interval. This data is available between 2013-07-30 and 2013-09-06. A handheld Garmin GPS eTrex Vista HCx was used in the field to mark injection and sampling locations. Further, an automatic weather station (AWS) was installed on Rabots glaciär, which stores meteorological data every 15 minutes. Pia Eriksson’s
thesis deals with this data. For the purpose of this work, mean hourly temperature (1 m above the surface) and total hourly precipitation data was extracted.

4.2 Data Processing

4.2.1 The Software FLUO

The software FLUO that comes with the GGUN-FL30 instrument allows for reading and processing of data from the data logger. Calibration is performed in the software according to each type of dye used, as the volumetric mass of Uranine 33.3% and RWT 20% differs and needs to be adjusted accordingly. If several tracers were logged simultaneously, data processing must be executed separately for each tracer. FLUO also enables the calculation of discharge by means of breakthrough curve analysis. This was done for each of the river dye tracing experiments. A mean of five runs of each experiment was used to reduce user error in defining the start and end of the breakthrough curve. Every breakthrough curve was calibrated and analysed separately to avoid inaccuracies.

4.2.2 Breakthrough curve shapes

The shape of the dye return curve may refer to various drainage configurations. Common features that characterize a curve and may even occur within the same curve are listed here. They have been reported in various studies (e.g. Seaberg et al. 1988; Willis et al. 1990)

- A distinct, well-defined, single peak.
- Multiple peaks, when more than one distinct peak occurs.
- Secondary peaks, when another, minor distinct peak occurs on the rising or falling limbs.
- A noisy pattern with several random short-lived peaks.
- A tail, when the curve is positively skewed showing a long tail to the right.
- The width of the peak, ranging from broad to narrow.

Since the visual interpretation of dye return curve shapes is very subjective, a number of calculations are carried out to describe characteristics of the dye return and ensure comparability between curves. The procedure is examined in the following section.
4.2.3 Dye tracing calculations

The GPS points of each dye injection location were imported into ArcGIS and converted into the common Swedish reference system SWEREF99 TM. The calculations of the dye tracing experiments were performed following Seaberg et al. (1988) and Willis et al. (1990; 2012).

a. A straight horizontal line between injection point and sampling location defines the transit distance \( x \) [m]. ArcGIS was used for the distance measurements. After adding the coordinates, injection and detection points were merged and connected with a polyline. The function ‘calculate geometry’ allows the computation of the distance. Due to the different positions of the detection instrument FL30 in July and August, a correction of the transit distance was necessary. All experiments are adjusted to the FL30 July position, because it is closest to the glacier snout. The stream distance is subtracted as follows:

\[
x_{\text{corrected}} = x - x_{\text{stream}}
\]  

b. Residence time \( t_m \) [s] defines the duration between dye injection and peak concentration. This value also needs to be corrected for the travel time in the stream. The injection location of discharge experiment D2 was the same as the FL30 July location. Therefore, direct values for discharge \( Q \) and travel time were available for the stream distance \( x_{\text{stream}} \). Based on this relation and available discharge data, the travel times for the remaining August experiments could be derived.

\[
t_{\text{corrected}} = t - t_{\text{stream}}
\]  

c. An estimation of the mean throughflow velocity is given by the transit speed \( v \) [m s\(^{-1}\)] for July using the following equation:

\[
v = \frac{x}{t_m}
\]

Corrected for the August experiments:

\[
v_{\text{corrected}} = \frac{x_{\text{corrected}}}{t_{\text{corrected}}}
\]
The dispersion coefficient $D \text{[m}^2\text{s}^{-1}]$ is a model to estimate the width of the breakthrough curve. It therefore indicates how the dye spreads out in the system. Equation (5) shows the calculation of $D$, where the corrected values for $x$ and $t_m$ are used for the August experiments. The variable $t_j$ represents the time [s] until half the peak concentration on the rising ($t_j = t_1$) and falling limb ($t_j = t_2$) is reached.

\[
D = \frac{x^2(t_m - t_j)^2}{4 t_m^2 t_j \ln \left( \frac{2}{t_m / t_j} \right)}
\]

The equation is first solved for $t_j = t_1$ to obtain $D_1$ and then for $t_j = t_2$ to get $D_2$. Both equations are solved simultaneously by changing the value of $t_m$ until the condition $D_1 = D_2$ is fulfilled.

e. Based on the dispersion coefficient the concentration-time curves are modelled by means of equation (6).

\[
c(t) = \frac{v}{Q} \frac{V_0}{\sqrt{(4\pi Dt)}} \exp \left( -\frac{(x - vt)^2}{4Dt} \right)
\]

In this equation $c(t)$ is the concentration of dye at time $t$ and $V_0$ represents the volume of injected dye. The discharge is defined as $Q$ and may be modified when fitting the modelled breakthrough curve to the measured one. The remaining variables were stated previously. For the modelled breakthrough curves, a time interval of 30 s was used to ensure comparability between modelled and measured curves.

Temporary storage of dye causes the falling limb of the dye return curve to be more elongated. The parameter storage retardation (SR) quantifies the percentage area difference under the measured and modelled falling limb and has been reported in a view studies (e.g. Hock and Hooke 1993; Nienow 1993; Schuler 2002; Cowton et al. 2013). The higher the SR, the lower the fit between modelled and measured curve, thus the higher the temporary storage.
g. Dispersivity \( d \) [m] describes the spreading rate of dye relative to the transit velocity. According to Seaberg et al. (1988) and Willis et al. (1990) it reveals the complexity of the transit route. It is given by:

\[
d = \frac{D}{v}
\]

(7)

h. The mass of dye, which passed through the detection location during the experiment is termed dye recovery \( W \) [g]. It consists of the tracer concentration \( c \), the discharge at the detection site \( Q \) and the time interval \( dt \).

\[
W = \sum_{t=1}^{n} c Q dt
\]

(8)

i. The percentage dye recovery is computed by using equation (9), where \( W_0 \) is the initial mass of dye that was injected. Caution needs to be exercised with regard to the conversion of dye volume to mass of the specific dye used. It is different for Uranine 33.3\% and RWT 20\%.

\[
W\% = 100 \frac{W}{W_0}
\]

(9)

The computation of dye recovery was not done for the July experiments, because no discharge recordings were available at that point.

### 4.2.4 Creating a Rating Curve

By definition, a rating curve represents the connection between two stream variables, such as discharge [m\(^3\) s\(^{-1}\)] and related water level or water pressure. They are used for the creation of stream outlet hydrographs. The principal idea is to infer a variable that is difficult to monitor continuously (discharge) from another related variable that is easier to measure (e.g. water level). To estimating discharge by means of dye tracing experiments, also termed dilution gauging, the distance between injection and detection needs to be long enough to ensure full mixing of the initial solution (Willis 2009; Hubbard and Glasser 2005). Five dye tracing experiments were conducted in the proglacial stream of Rabots glaciär and flow rates were calculated for the
particular times, representing both high and low flow conditions were captured. The flow rates were computed with the software FLUO.

A suitable position for monitoring stage is a stable stream cross section with a high depth-to-width ratio. In the proglacial stream of Rabots glaciär the instrument was placed at a stable section after the braided systems. Since the gauging station instrument measures the distance from the sensor to the water surface, the values needed to be subtracted from an arbitrary value to obtain the relative water height. For this purpose a value of 2 m was used. The calculated discharge values were related to the respective gauging height. Therefore, an average of stage for the duration of the respective dye tracing experiment was used. A logarithmic trendline was fitted through the known stage-discharge points. Based on the formula and the continuous stage measurements, the remaining discharge values were computed.

The rating curve based on compensated water pressure was derived in a similar manner. However, the data preparation required a different handling. The instrument measured the absolute pressure, i.e. water pressure and atmospheric pressure. To obtain the fluctuations in water pressure, the atmospheric pressure was subtracted from the water pressure. Since the instrument was adjusting to the environment, it did not provide precise data during the first hours. Those values have been removed, because they are not representative for the actual conditions. The final processing of the rating curve followed that used for stage.

4.2.5 Discharge Data Analyses

For comparison of both independent rating curves, the exact time steps had to be examined. As the stage recordings were done every 15 minutes and the water pressure measurements every 10 minutes, the overlapping time step is every half-hour. The data was prepared accordingly. In order to make interpretations of the catchment hydrology, diurnal cycles for the whole season and at daily or weekly levels were examined. Moreover, calculations (presented previously in section 4.2.3) were performed as well as the computation of the model efficiency. When analysing the hydrographs, a mean of the water pressure and stage-derived discharge calculations was used. However, the water pressure recording was available only from 2013-08-30, which is why just the stage-derived discharge data was used for the previous time period (starting 2013-08-22).
4.2.6 Field Turbidity Data

During some dye tracing experiments, turbidity data was recorded by the FL30 instrument (see table 4-1 for details). It is a measure of cloudiness in the water expressed as the absolute unit NTU (nephelometric turbidity units). High turbidity values are caused by sediments in the water, bacteria and germs or chemical precipitates. Hubbard and Glasser (2005) highlight that suspended sediments in particular reduce the transparency of glacial meltwater. They further point out that it is hard to calibrate turbidity signals in terms of suspended sediment concentration (SSC), because a number of properties (e.g. grain-size distribution, shape of particles, lithology etc.) influence the measured values. Thus, turbidity signals are usually reported as measured or relative turbidity.

Since no calibration standard was available for our study, the turbidity data is reported as measured values. From the available data, the diurnal cycle was retrieved as well as the correlation to discharge. Statistical analyses, including the coefficient of variation (section 4.2.9) were performed in order to explain the results quantitatively.

4.2.7 Digital Elevation Models (DEMs)

There were several digital elevation models available for our study area. The Swedish national mapping agency Lantmäteriet derived a DEM from aerial photography in 2008. The geometric resolution is 10m. It was used for modelling the hillshades.

Keith Brugger provided another DEM dating back to 2011, which was created through mapping in the field with differential GPS. The geometric resolution is 10 m. As it is just two years old, it is assumed to be more representative for the current glacier extent. It covers the whole glacier area.

To cover the whole catchment, an ASTER global digital elevation model (GDEM) was retrieved from the USGS EarthExplorer. It has an average spatial resolution of 1 arc-second, which corresponds to about 30 m. When working with the data it turned out that the cell size was about 20m. The latest ASTER GDEM Version 2, released in 2011, was used. According to the ASTER GDEM Validation Team (https://lpdaacaster.cr.usgs.gov/GDEM/Summary_GDEM2_validation_report_final.pdf, 17 Jan., 2014) the overall accuracy is approximately 17 m at a 90% confidence level.
However, a large number of undulation effects were registered when evaluating the vertical resolution (e.g. Suwandana et al. 2012).

The DEMs were used for different purposes. Based on the mapped glacier extent and proglacial stream (see following section for details), an outline map was produced. Therefore, all available data had to be projected into the Swedish coordinate system SWEREF TM99 (Project / Data Management toolbox). The DEM provided by Keith Brugger was used for this map, as it seemed most accurate. First, the DEM was cut to the glacier extent (Clip / Analysis toolbox). Then, contour lines were derived from the surface (Contour / Spatial analyst toolbox) at different intervals.

The tool *hillshade* (Spatial Analyst toolbox) produces a shaded relief by including illumination source angle and shadows on basis of a DEM. The source is thereby set infinite, i.e. no weather conditions (cloudiness) are considered. The local horizon is considered at each cell and cells in shadow receive a zero value. If the option *model shadows* is enabled, impacts of local illumination angle and shadow are included. The National Oceanic and Atmospheric Administration (NOAA) provides a Solar Position Calculator (http://www.esrl.noaa.gov/gmd/grad/solcalc/azel.html, 15 Jan., 2014). By entering the latitude/longitude coordinates and the date/time of interest, solar declination, azimuth and elevation are obtained. In this work, the shadow extent at time of the maximum melt was obtained for the 15th of each month, i.e. May until September. Since the melt follows the diurnal temperature pattern (e.g. Benn and Evans 2010), the daily maximum temperature was extracted from the temperature recordings for each month. The 2008 DEM (Lantmäteriet) was used for hillshade analysis, because of its good geometric resolution. The shaded relief raster has values from 0 (shaded) to 255 (no shadow). Based on visual interpretations, a threshold of 50 was applied to extract the shaded areas as a layer.

For delineating the catchment, several tools from the Spatial Analyst toolbox/Hydrology toolset were used. There may be imperfections (sinks) within the DEM, i.e. cells without defined drainage value. However, they are important in order to determine the flow direction out of the cell later on. Therefore, the *Fill* tool was used prior to the other work steps. Next, the flow direction for each cell was derived using the *Flow Direction* tool. This step is crucial as it reveals how the water is flowing across the landscape surface. The *Flow Accumulation* tool accumulates the cells flowing into a particular downslope cell. Every cell within the grid then contains information
on the number of cells upstream of that respective cell. Thus, cells with a high flow accumulation are situated in regions of lower elevation, i.e. in valleys or drainage channels. A Pour Point is defined as location where water flows out of the cell. The user may set it arbitrary – in this case it is set to the position of the gauging station. By means of the Snap Pour Point tool, the set point is snapped to the most proximate area of high flow accumulation and converted to the raster format. The function Watershed finally delineates the contributing area to the previously defined pour point by including the result of the flow direction analysis, allowing for delineation of drainage divides. For this purpose the ASTER 2011 DEM was used, because it covered the whole Kebnekaise area and not just sections of the catchment.

4.2.8 Information from Landsat 8 Imagery

In order to map the extent of Rabots glaciär and the proglacial stream, several Landsat 8 images were retrieved from the USGS EarthExplorer (Path 195; Row 13). They were evaluated visually regarding cloudiness around the study area. It turned out that the image required on 2013-08-31 was free of clouds and suitable for mapping the extent. Prior to mapping the image was processed with the image classification software ENVI. To increase the geometric resolution pansharpening was conducted. Thus, the resolution increased from 30 m to 15 m. The mapping was done in ArcMap using the Editor and different band combinations to separate the glacier area from different surfaces. Following band combinations were used:

- Natural Colour: bands 4, 3, 2
- Colour Infrared (CIR): bands 5, 4, 3
- Shortwave Infrared (SIR): bands 7, 5, 4

Further, the Normalized-Difference Snow Index (NDSI) helped to distinguish snow and glacier ice from other surfaces (e.g. Hall and Riggs 2011). It is computed by using the Landsat 8 channels Green (band 3) and SWIR 1 (band 6) as follows:

\[
NDSI = \frac{\text{Band3} - \text{Band6}}{\text{Band3} + \text{Band6}} \tag{10}
\]
4.2.9 Statistical analyses

Besides commonly used statistical variables such as minimum, maximum, mean or standard deviation, further statistical analyses were conducted. They are explained briefly in this section.

a. The coefficient of variation (CV) is a non-dimensional value defined as the ratio of the standard deviation (sd) to the arithmetic mean.

\[
CV = \frac{sd}{mean} \tag{11}
\]

b. The coefficient of determination (R\(^2\)) specifies the degree of linear correlation of variables, for example simulated and observed data. It defines the part of variance in the data that is explained by the regression model. It is calculated by squaring Pearson’s correlation coefficient r and ranges from 0 to 1, with higher values showing less error variance. Values greater than 0.5 are commonly regarded to be acceptable (e.g. Santhi et al. 2001). In this case O and P represent observed and predicted values.

\[
R^2 = \left(\frac{\sum(O_i - \bar{O})(P_i - \bar{P})}{\sqrt{\sum(O_i - \bar{O})^2 \sum(P_i - \bar{P})^2}}\right)^2 \tag{12}
\]

c. The Nash-Sutcliffe efficiency (E) is a normalized measure of the noise (remaining variance) to signal (measured data variance) ratio. The dimensionless value indicates how well two datasets (e.g. observed vs. simulated) fit 1:1. The range is between minus infinity and 1 with 1 being the optimal value. Acceptable values are considered to be above 0 (Nash and Sutcliffe 1980).

\[
E = 1 - \frac{\sum(O_i - P_i)^2}{\sum(O_i - \bar{O})^2} \tag{13}
\]

d. The index of agreement (IOA) by Willmott (1981) was developed in order to mitigate the problem for over- and underprediction as faced with R\(^2\) or E. It is dimensionless and determined by following equation:
The root mean square error (RMSE) is a standard variable in error statistics and calculated as follows in equation 14. Generally, the smaller the RMSE, the better the model performance.

\[
RMSE = \sqrt{\frac{\sum (O_i - P_i)^2}{\sum (|P_i - \bar{O}| + |O_i - \bar{O}|)^2}}
\]  

f. Singh et al. (2004) developed a standardized approach to specify what a low RMSE means. The so-called RMSE-observations standard deviation ratio (RSR) is obtained from the ratio of the RMSE and the standard deviation of observed data (sd_{obs}). It varies from the optimal value of 0 (i.e. perfect model performance) to large positive values.

\[
RSR = \frac{RMSE}{sd_{obs}}
\]  
g. Percent bias (PBIAS) is a measure of tendency of the predicted data to be smaller or larger than the observed counterparts. Low magnitude values indicate accurate model performance with 0 being the optimal value. Positive values show an underestimation bias whereas negative results indicate overestimation bias. It is calculated by means of equation 16.

\[
PBIAS = \left[ \frac{\sum (O_i - P_i) * 100}{\sum (O_i)} \right]
\]


5 RESULTS

5.1 Field Observations

Field observations from 2013-07-04 are summarized in a report by Susanne Ingvander (pers. comm., 22 Jul., 2013). Due to the early onset of melt in the end of May/beginning of June respectively, almost no snow was left on and around the glacier, furthermore, little liquid water was observed on the glacier. Open moulins were recorded only on the far north side. In late July (2013-07-22) another short dye tracing investigation was done. However, no further information on general observations are available.

During our field campaign one month later (starting 2013-08-08), no snow remained off-glacier. While conducting manual sampling close to the glacier front on an island between both main outlet streams (on 2013-08-15 till 2013-08-16), a rising water level was noted during late afternoon. It was highest from around 5 pm until 7 pm, falling gradually after that. It was also observed that the water tends to spread out into the braided systems with rising water level.

![Fig. 5-1. Proglacial stream of Rabots glaciär. The three braided systems (BS) are visible, although the third one extends further downstream. The two main outlets, termed RJA (north) and RJB (south) are depicted on the right (left photograph by Per Holmlund, right photographs by Caroline Coch)](image_url)
The photograph in Fig. 5-1 shows the proglacial system with the three braided systems (abbreviated BS). The difference in sediment load in both main outlet streams, named RJA and RJB, was visible. The stream on the northern side appeared to carry much more sediments than the one to the south. Besides the main outlets a few very small ones emerge supraglacially from the glacier. Also an esker in flow direction was located approximately 150 m away from the glacier front.

5.2 Catchment area

The map in Fig. 5-2 illustrates the result of the watershed delineation. The catchment has a size of 9.42 km$^2$, of which 3.08 km$^2$ (i.e. 32.67%) are assigned to Rabots glaciär. The steepest slope is 79.55° with an average is 26.34°. The heights range from 1019 m a.s.l. to 2088 m a.s.l., with a mean of 1428 m a.s.l. It is visible that the watershed boundary follows the highest elevations. The gauging station (red point in the map) was used as the pour point and represents the outlet point for the catchment.

Fig. 5-2. Catchment around Rabots glaciär. The gauging station was used as pour point for deriving the watershed. Due to the 3D view the scale will deviate in the sloping areas (ASTER 2011 DEM, glacier extent based on Landsat 8 image acquired 2013-08-31).
5.2.1 Breakthrough curve properties

Table 5-1 summarizes the results of the five river dye tracing experiments conducted between 2013-08-08 and 2013-08-18 while Fig. 5-3 illustrates the respective breakthrough curves including measured and modeled concentrations as well as percentage dye recovery. As mentioned earlier, the injection and recording points for the river dye tracing experiments varied. All braided systems were included during experiments D3 and D5. For experiment D4 the braided systems were excluded, i.e. just a channelized section of the stream was used. One or two braided systems were involved in experiments D1 and D2. The transit distance $x$ varied accordingly; the shortest in experiment D4 (408 m) and the longest in experiment D5 (1534 m).

The shape of the dye return curves is very similar for all experiments, characterized by a narrow, well-defined distinct peak and just few (if any) noisy patterns on the falling limb (e.g. in experiment D2). Experiment D4 exhibits a very narrow curve, which is due to the short distance and channelized flow, whereas the remaining curves are broader relative to D4. The maximum concentrations recorded at the gauging station vary between 23.04 ppb (D2) and 106.92 ppb (D4) and reflect the amount that was originally injected.

The calculated discharge was highest on 2013-08-18 (2.36 m$^3$ s$^{-1}$), and lowest on 2013-08-10 (1.08 m$^3$ s$^{-1}$), while the throughflow velocity, shows a maximum of 0.58 m s$^{-1}$ (D5) and a minimum of 0.23 m s$^{-1}$ (D2).

Table 5-1. Overview of river dye tracing results including assigned code, date, braided systems included, transit distance $x$, type and amount of dye injected $W_0$, sampling method, local injection time, transit time $t_m$, value of peak concentration, throughflow velocity $v$, dispersion coefficient $D$, dispersivity $d$, storage retardation $SR$, percentage dye recovery $W\%$ and computed discharge $Q$. 

<table>
<thead>
<tr>
<th>Code</th>
<th>Date</th>
<th>Braided systems</th>
<th>$x$ [m]</th>
<th>Dye</th>
<th>$W_0$ [ppb]</th>
<th>Sampling method</th>
<th>Local injection time</th>
<th>Transit time $t_m$ [s]</th>
<th>Peak conc. [ppb]</th>
<th>$D$ [m$^2$]</th>
<th>$d$ [m]</th>
<th>$SR$ [%]</th>
<th>$W%$ [%]</th>
<th>$Q$ [m$^3$ s$^{-1}$]</th>
</tr>
</thead>
<tbody>
<tr>
<td>D1</td>
<td>2013-08-08</td>
<td>BS3</td>
<td>685</td>
<td></td>
<td>30</td>
<td>auto</td>
<td>12:37</td>
<td>1800</td>
<td>25.76</td>
<td>0.38</td>
<td>0.59</td>
<td>1.56</td>
<td>38.50</td>
<td>55.05</td>
</tr>
<tr>
<td>D2</td>
<td>2013-08-10</td>
<td>BS2, BS3</td>
<td>874</td>
<td></td>
<td>30</td>
<td>auto</td>
<td>10:52</td>
<td>3840</td>
<td>23.04</td>
<td>0.23</td>
<td>1.46</td>
<td>6.43</td>
<td>20.04</td>
<td>81.50</td>
</tr>
<tr>
<td>D3</td>
<td>2013-08-15</td>
<td>BS1, BS2, BS3</td>
<td>1237</td>
<td></td>
<td>60</td>
<td>auto</td>
<td>12:37</td>
<td>3450</td>
<td>45.25</td>
<td>0.36</td>
<td>2.91</td>
<td>8.11</td>
<td>24.77</td>
<td>93.52</td>
</tr>
<tr>
<td>D4</td>
<td>2013-08-18</td>
<td>None</td>
<td>408</td>
<td></td>
<td>50</td>
<td>auto</td>
<td>12:22</td>
<td>1710</td>
<td>196.92</td>
<td>0.24</td>
<td>0.10</td>
<td>0.41</td>
<td>17.90</td>
<td>77.69</td>
</tr>
<tr>
<td>D5</td>
<td>2013-08-18</td>
<td>BS1, BS2, BS3</td>
<td>1534</td>
<td></td>
<td>100</td>
<td>auto</td>
<td>13:26</td>
<td>2640</td>
<td>66.67</td>
<td>0.58</td>
<td>4.11</td>
<td>7.08</td>
<td>31.31</td>
<td>89.18</td>
</tr>
</tbody>
</table>
Fig. 5-3. Breakthrough curves, modeled curves and dye recovery retrieved from the river dye tracing experiments in August. The dye RWT 20% was used during all the experiments.
Table 5-2: Model efficiency statistics evaluating model performance for the five river dye tracing experiments. All model simulations show a very good performance.

<table>
<thead>
<tr>
<th></th>
<th>$R^2$</th>
<th>PBIAS [%]</th>
<th>RMSE</th>
<th>RSR</th>
<th>E</th>
<th>IOA</th>
</tr>
</thead>
<tbody>
<tr>
<td>D1</td>
<td>0.76</td>
<td>8.49</td>
<td>33.73</td>
<td>0.52</td>
<td>0.73</td>
<td>0.93</td>
</tr>
<tr>
<td>D2</td>
<td>0.92</td>
<td>0.08</td>
<td>0.09</td>
<td>0.00</td>
<td>1.00</td>
<td>1.00</td>
</tr>
<tr>
<td>D3</td>
<td>0.94</td>
<td>-0.15</td>
<td>0.98</td>
<td>0.01</td>
<td>1.00</td>
<td>1.00</td>
</tr>
<tr>
<td>D4</td>
<td>0.95</td>
<td>0.78</td>
<td>47.22</td>
<td>0.22</td>
<td>0.95</td>
<td>0.99</td>
</tr>
<tr>
<td>D5</td>
<td>0.87</td>
<td>7.13</td>
<td>84.08</td>
<td>0.36</td>
<td>0.87</td>
<td>0.97</td>
</tr>
</tbody>
</table>

Values for the dispersion coefficient range between 0.10 m$^2$ s$^{-1}$ (D4) and 4.11 m$^2$ s$^{-1}$ (D5). They coincide with the dimensions of the dispersivity, with 0.41 m being the lowest (D4) and 4.11 m the greatest (D5). For all experiments, the percentage dye recovery amounts above 55%, the highest being 93.5% (D3). Storage retardation ranges below 38.5% (D1) with a minimum of 17.8% (D4).

Generally, the model simulates the observed values quite well as shown by the model efficiency statistics in table 5-2. All statistical measures show a very good agreement between simulated and observed data. The values for $R^2$ amount at least 0.76 during the experiments, E is generally above 0.73 and IOA greater than 0.93. PBIAS lies below 8.49%, tending to underestimate the measured data except of experiment D3. Also RMSE and RSR show a little error tendency in their values. The best model performance was achieved during experiment D2 showing a zero RSR with E and IOA being 1.

5.2.2 Rating curves

The plots in Fig. 5-4 show the obtained rating curves from relative stage or water pressure as well as the computed (interpolated) discharge points for the season. It is visible that the majority of computed discharge points are situated outside the regression lines. The formula for the logarithmic trendline is given in the diagrams, as well as the coefficient of determination $R^2$. It reveals how well the known discharge data points fit the logarithmic curve. It amounts 0.89 for the stage derivation and 0.86 for the water pressure derivation.
5.2.3 Discharge plots

The plot (Fig. 5-5) shows the discharge for the monitoring season. The coefficient of variation included all time steps and amounts to 0.38 for the entire period with minimum and maximum discharges of $0.32 \, \text{m}^3\,\text{s}^{-1}$ (2013-07-31 16:00) and $3.92 \, \text{m}^3\,\text{s}^{-1}$ (2013-07-31 16:00). An individual consideration of monthly and weekly variability is presented in table 5-3.

The coefficient of variation is shown to be highest during August (0.37) compared to the available values from July (0.22) or September (0.26). However, the mean discharge is highest in July ($2.37 \, \text{m}^3\,\text{s}^{-1}$) and gradually decreasing during August ($1.72 \, \text{m}^3\,\text{s}^{-1}$) and September ($1.24 \, \text{m}^3\,\text{s}^{-1}$). When looking at the particular weeks it is clear that variability increases between weeks 31 and 35.
RESULTS

Table 5-3. Statistical evaluation of discharge variations for the whole monitoring season, each month and each calendar week respectively. Note that the monitoring period was from 2013-07-22 until 2013-09-06, i.e. July, September and week 36 are not fully covered.

<table>
<thead>
<tr>
<th></th>
<th>All values</th>
<th>July</th>
<th>August</th>
<th>September</th>
<th>Week 30</th>
<th>Week 31</th>
<th>Week 32</th>
<th>Week 33</th>
<th>Week 34</th>
<th>Week 35</th>
<th>Week 36</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Q_{\text{max}}$ [m$^3$s$^{-1}$]</td>
<td>0.32</td>
<td>1.34</td>
<td>0.32</td>
<td>0.57</td>
<td>1.34</td>
<td>2.4</td>
<td>0.9</td>
<td>1.05</td>
<td>0.64</td>
<td>0.32</td>
<td>0.75</td>
</tr>
<tr>
<td>$Q_{\text{mean}}$ [m$^3$s$^{-1}$]</td>
<td>3.92</td>
<td>3.92</td>
<td>3.36</td>
<td>1.89</td>
<td>3.66</td>
<td>3.92</td>
<td>2.9</td>
<td>3.26</td>
<td>2.86</td>
<td>2.85</td>
<td>1.89</td>
</tr>
<tr>
<td>$Q_{\text{mean}}$ [m$^3$s$^{-1}$]</td>
<td>1.79</td>
<td>2.37</td>
<td>1.72</td>
<td>1.24</td>
<td>2.14</td>
<td>2.87</td>
<td>1.78</td>
<td>1.65</td>
<td>1.46</td>
<td>1.49</td>
<td>1.28</td>
</tr>
<tr>
<td>CV$^*$</td>
<td>0.38</td>
<td>0.22</td>
<td>0.37</td>
<td>0.26</td>
<td>0.18</td>
<td>0.15</td>
<td>0.28</td>
<td>0.35</td>
<td>0.38</td>
<td>0.49</td>
<td>0.25</td>
</tr>
</tbody>
</table>

* The coefficient of variation considers the discharge values of every time step, i.e. every half-hour.

up to a maximum of 0.49. The mean diurnal pattern of discharge is illustrated for the monitoring season and for each month and week in Fig. 5-6. The average daily pattern for the entire monitoring season shows a rising limb from 8:00, reaching a maximum between 16:00 and 18:00 and a falling limb subsequently. The peakflow period appears to be a little noisy, showing two small peaks at 16:00 and 18:00. When looking at the differences between each month, the peaks in July are more distinct than the entire monitoring season, occurring at 15:00 and 18:00. In August, the peaks are not as distinct. One occurs between 16:00 and 16:30 and the other one at 17:30. The pattern for September no longer follows a typical diurnal cycle. There are two distinct peaks, one at 15:00 and the other one at 17:00. The limb starts to rise at 8:00 until reaching the second peak, then drops until 19:30 and starts to rise thereafter.

The peak timing for each week reveals a very dynamic pattern. The peaks are most distinct in the first monitoring week (Nr. 30), appearing at 15:00 and 18:00. The next week (Nr. 31) shows a quite noisy pattern of multiple peaks between 12:00 and 14:00 as well as between 16:00 and 17:30. Compared to the previous week, the timing is earlier and the peaks are not as distinct. Week 32 is also characterized by several smaller peaks, appearing 12:00 and between 16:30 and 18:30. The subsequent week (Nr. 33) shows a smoother pattern with small peaks at 15:00 and between 17:30 and 18:30. The fluctuations seem to increase again with week 34, showing a number of peaks between 12:30 and 18:00. Almost no diurnal pattern is visible in week 35. Then again, week 36 shows a rising limb with a very distinct peak at 15:00, followed by a falling limb.

To summarize, the two distinct peaks from the first week tend to occur earlier in the day and to even out as the season progresses. Reasons for that are going to be discussed in a later chapter.
Fig. 5-6. Mean diurnal runoff cycles for the monitoring season (top left), each month (down left) and each week (right). The weekly graphs show the actual data in the background (dotted lines) and a moving average of two hours (solid line).
5.3 Glacier Dye tracing

5.3.1 Breakthrough curve properties

The measured and modeled breakthrough curves as well as the calculated dye recovery (where applicable) for all glacier dye tracing experiments are illustrated in Fig. 5-7. The model performance is evaluated in table 5-4.

The form of the recorded curves varies significantly between experiments. The July experiments show generally a noisy pattern with several random short-lived peaks. The curve in experiment J1 shows a steep rising limb reaching the broad peak of 7.4 ppb after approximately 1 h 30 min, followed by a long tail that does not go back to baseline concentration during the recording time of 4 hours. On the falling limb, after around 2 h 30 min, there is a minor secondary peak looking like a plateau before the curve is gradually falling. The model is not able to capture the dynamics as revealed by visual interpretations as well as the efficiency statistics. \( R^2 \) is below 0.5 and E is even negative. 70.82% is estimated to be stored in the system.

<table>
<thead>
<tr>
<th></th>
<th>( R^2 )</th>
<th>PBIAS [%]</th>
<th>RMSE</th>
<th>RSR</th>
<th>E</th>
<th>IOA</th>
</tr>
</thead>
<tbody>
<tr>
<td>J1</td>
<td>0.33</td>
<td>38.25</td>
<td>57.28</td>
<td>1.04</td>
<td>-0.08</td>
<td>0.72</td>
</tr>
<tr>
<td>J2</td>
<td>0.48</td>
<td>34.42</td>
<td>8.11</td>
<td>0.84</td>
<td>0.29</td>
<td>0.79</td>
</tr>
<tr>
<td>J3</td>
<td>0.77</td>
<td>18.26</td>
<td>210.91</td>
<td>0.46</td>
<td>0.78</td>
<td>0.95</td>
</tr>
<tr>
<td>A1</td>
<td>0.98</td>
<td>2.97</td>
<td>26.11</td>
<td>0.13</td>
<td>0.98</td>
<td>1.00</td>
</tr>
<tr>
<td>A2</td>
<td>0.95</td>
<td>-1.61</td>
<td>9.13</td>
<td>0.22</td>
<td>0.95</td>
<td>0.99</td>
</tr>
<tr>
<td>A3</td>
<td>0.97</td>
<td>1.10</td>
<td>38.83</td>
<td>0.19</td>
<td>0.96</td>
<td>0.99</td>
</tr>
<tr>
<td>A4</td>
<td>0.96</td>
<td>-1.33</td>
<td>5.64</td>
<td>0.20</td>
<td>0.96</td>
<td>0.99</td>
</tr>
<tr>
<td>A7</td>
<td>0.98</td>
<td>-4.45</td>
<td>6.94</td>
<td>0.16</td>
<td>0.97</td>
<td>0.99</td>
</tr>
</tbody>
</table>
Fig. 5-7. Breakthrough curves of the dye tracing experiments in July and August. Note that modeled breakthrough curves could be calculated only for the August experiments since discharge was not recorded during the earlier experiments.
Experiment J2 shows a very weak return signal with a peak of 1.43 ppb after 2 h 40 min. The entire breakthrough curve exhibits a noisy pattern with many random peaks both on the rising and on the falling limbs. The tail is very long and does not reach the base level during the experiment time of 6 hours. Although an improvement on J1, the model also cannot capture the form of the return curve for experiment J2. E is in the positive range and RSR and IOA reach quite high values. However, $R^2$ is still below 0.5. Storage retardation is approximately 58.51%.

The third breakthrough curve in July (experiment J3) has a distinct, well-defined peak of 52.89 ppb around 2 h 15 min since injection, with a slightly noisy pattern. The falling limb does not reach the baseline concentration until 4 h 15 min but its shape suggests a long tail to the right. The model could capture the breakthrough curves dynamics as indicated by acceptable values for $R^2$, E and IOA. The amount of storage retardation is approximately 43.86%.

The dye return curve of experiment A1 shows a relatively broad peak of 45.22 ppb after approximately 45 min. The rising limb appears as a long tail. The percentage dye recovery was computed to be 70.7% whereas storage retardation was 10.71%. The model shows a very good fit with $R^2$, E and IOA close to 1, and very small PBIAS and RMSE and RSR respectively.

The breakthrough curve of experiment A2 appears to have a small, narrow peak of 9.49 ppb after approximately 1 h and another, broader peak right after. The tail is relatively short, showing a slightly noisy pattern. The dye recovery ranges below 20%, storage retardation below 13.84%. According to the negative value of PBIAS, the model has a slight overestimation bias. Generally, there is a very good agreement between modeled and measured data as proved by the efficiency statistics.

The resulting dye return curve in experiment A3 shows a distinct, well-defined peak of 29.47 ppb after around 3 h 30 min with little noisy behavior. Of all glacier dye tracing experiments, the rate of dye recovery was highest during this experiment (78.69%) and storage retardation quite low (10.78%). The model performance is considerably very good as indicated by the statistics.

The recorded curve of experiment A4 has the lowest residence time (35 min), showing several peaks, the greatest amounting 7.96 ppb. The short-lived peaks continue on the falling limb. The percentage dye recovery is very low (8.77%) such as storage retardation (11.03%). As shown by the model evaluation statistics, the performance is quite good with a small overestimation bias.
Experiment A7 was recorded using both, automatic and manual sampling (see Fig. 5-8). The manual sampling was conducted in the two main outlet streams, abbreviated as RJA (northern one) and RJB (southern one). The automatically recorded breakthrough curve shows a steep rising limb with a peak of 1.54 ppb after around 15 hours and increased noise on the falling limb. Dye recovery amounts to just 8.25%, thus showing the lowest dye recovery value of all the glacier dye tracing experiments. The model shows a very good agreement between simulated and observed values. The manual sampling in stream RJA did not show a signal of dye return but a very noisy pattern, perhaps due to turbidity. Since the pattern already occurred even before the dye injection, the curve can be disregarded. In contrast, the manual sampling in the southern stream (RJB) showed a dye return curve with a peak of 10.78 ppb approximately 15 h 50 min after the injection. It is generally characterized by a noisy pattern, most likely caused by the instrument’s sensitivity to turbidity. The value of the dye recovery was computed to be much higher than for the automatic sampling, which is 23.70%. The same applies to the value of storage retardation, which amounts 58.56%, compared to 5.03% of the automatic sampling. The model performance is not as good as in the previous experiments, but still acceptable with $R^2$ above 0.5, a small overestimation bias and high values for E and IOA. A closer comparison between the results of automatic and manual sampling will follow in discussion section 6.3.7.

---

Fig. 5-8. Results of the manual sampling in both main outlet streams during experiment A7. The signal of RJA is regarded as noise, since it already occurred before the dye was injected and no distinct signal could be identified. The dotted line represents the model result of curve RJB.
5.3.2 Dye Tracing Calculations

The summary table below (Table 5-5) lists the results of the dye tracing calculations, including also the result of the manual sampling in stream RJB. The transit distance is ranging between 488 m in experiment A4 and 2339 m in experiment A7. The injection of the tracer was performed between early noon and late afternoon. The earliest injection took place at 11:27 during experiment J2, whereas the latest injection time was 16:46 (experiment A2). The residence time varied according to the transit distance, ranging between 2130 s (A4) and 57000 s (A7_manual). The values for the recorded peak concentration also show a big range, with 1.43 ppb being the lowest (J2) and 52.89 ppb being the highest (J3). The values of transit velocity are between 0.04 m s$^{-1}$ (A7) and 0.28 m s$^{-1}$ (J1).

The values for the dispersion coefficient show a great range with 0.19 m$^{2}$ s$^{-1}$ in experiment A3 and 24.27 m$^{2}$ s$^{-1}$ in experiment J1. Also the dispersivity varies significantly from 2.89 m (A3) to 92.63 m (J2). It should be noted that the results of automatic recordings and manual sampling in stream RJB differ greatly. The analyses of the manual sampling show higher values for residence time, peak concentration, dispersion coefficient, dispersivity as well as percentage dye recovery. Reasons for this are going to be discussed later on.

Table 5-5. Results of the successful glacier dye tracing experiments showing assigned code, date, injection location, transit distance, dye type and volume, sampling method, local injection time, time to half the peak concentration, peak concentration, throughflow velocity, dispersion coefficient, dispersivity, storage retardation and dye recovery.

<table>
<thead>
<tr>
<th>Code</th>
<th>Date</th>
<th>Injection Location</th>
<th>x [m]</th>
<th>Dye</th>
<th>W$_d$ [ml]</th>
<th>Sampling method</th>
<th>Local injection</th>
<th>$t_{50}$ [s]</th>
<th>Peak conc. [ppb]</th>
<th>v [m/s]</th>
<th>D [m$^2$/s]</th>
<th>d [m]</th>
<th>SR [%]</th>
<th>W [%]</th>
</tr>
</thead>
<tbody>
<tr>
<td>J1</td>
<td>2013-07-04</td>
<td>supraglacially</td>
<td>1489</td>
<td>RWT 20%</td>
<td>125</td>
<td>auto</td>
<td>14:47</td>
<td>5220</td>
<td>7.4</td>
<td>0.28</td>
<td>24.27</td>
<td>86.24</td>
<td>70.82</td>
<td>n.a.</td>
</tr>
<tr>
<td>J2</td>
<td>2013-07-05</td>
<td>Crevasse</td>
<td>2022</td>
<td>RWT 20%</td>
<td>245</td>
<td>auto</td>
<td>11:37</td>
<td>9690</td>
<td>1.43</td>
<td>0.21</td>
<td>19.33</td>
<td>92.63</td>
<td>58.51</td>
<td>n.a.</td>
</tr>
<tr>
<td>J3</td>
<td>2013-07-22</td>
<td>Moulin</td>
<td>1165</td>
<td>RWT 20%</td>
<td>250</td>
<td>auto</td>
<td>12:05</td>
<td>8013</td>
<td>52.89</td>
<td>0.15</td>
<td>4.04</td>
<td>27.79</td>
<td>43.86</td>
<td>n.a.</td>
</tr>
<tr>
<td>A1</td>
<td>2013-08-08</td>
<td>Moulin</td>
<td>655</td>
<td>RWT 20%</td>
<td>150</td>
<td>auto</td>
<td>16:33</td>
<td>2670</td>
<td>45.22</td>
<td>0.25</td>
<td>2.84</td>
<td>11.59</td>
<td>10.71</td>
<td>70.70</td>
</tr>
<tr>
<td>A2</td>
<td>2013-08-08</td>
<td>Moulin</td>
<td>703</td>
<td>Uranine 33.3%</td>
<td>100</td>
<td>auto</td>
<td>16:46</td>
<td>3690</td>
<td>9.49</td>
<td>0.19</td>
<td>0.81</td>
<td>4.25</td>
<td>13.84</td>
<td>18.79</td>
</tr>
<tr>
<td>A3</td>
<td>2013-08-09</td>
<td>Moulin</td>
<td>829</td>
<td>RWT 20%</td>
<td>150</td>
<td>auto</td>
<td>14:30</td>
<td>12900</td>
<td>29.47</td>
<td>0.06</td>
<td>0.19</td>
<td>2.86</td>
<td>10.78</td>
<td>78.69</td>
</tr>
<tr>
<td>A4</td>
<td>2013-08-09</td>
<td>Moulin</td>
<td>488</td>
<td>Uranine 33.3%</td>
<td>100</td>
<td>auto</td>
<td>15:29</td>
<td>2103</td>
<td>7.96</td>
<td>0.23</td>
<td>1.22</td>
<td>5.34</td>
<td>11.03</td>
<td>8.77</td>
</tr>
<tr>
<td>A7</td>
<td>2013-08-15</td>
<td>supraglacially</td>
<td>2339</td>
<td>RWT 20%</td>
<td>500</td>
<td>manual*</td>
<td>13:20</td>
<td>53640</td>
<td>2.54</td>
<td>0.04</td>
<td>0.79</td>
<td>18.01</td>
<td>5.03</td>
<td>8.25</td>
</tr>
</tbody>
</table>

*manual in the southern proglacial stream (abbreviated RJB)
5.4 Field Turbidity

The turbidity recordings in the proglacial stream of Rabots glaciär reflect the conditions within a period of 9 days in August. Therefore, the analyses focused on the diurnal cycle of turbidity and discharge. Table 5-6 shows summary statistics for each day of measurement and the mean of all days. Figure 5-9 illustrates the relationship between the diurnal cycles of turbidity and discharge.

As visible in the statistics, the correlation between turbidity and discharge is relatively poor except for 2013-08-16. However, looking at the diurnal cycle, the correlation increases with r of 0.67 and R² of 0.45. The diurnal turbidity cycle shows a generally noisy pattern and a distinct peak at 15:30. It approximately follows the pattern of the diurnal discharge cycle. The relative magnitude of the standard deviation was computed for both discharge and turbidity by means of the coefficient of variation. The results show that the relative variability is quite similar for all the days of measurement.

Table 5-6. Comparison between turbidity and discharge for each day of turbidity recording. Mean, minimum, maximum, coefficient of variation, Pearson’s correlation coefficient and coefficient of determination are given.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Q [m³ s⁻¹]</td>
<td>Q [m³ s⁻¹]</td>
<td>Q [m³ s⁻¹]</td>
<td>Q [m³ s⁻¹]</td>
<td>Q [m³ s⁻¹]</td>
<td>Q [m³ s⁻¹]</td>
</tr>
<tr>
<td>Min</td>
<td>152.06 1.69</td>
<td>152.06 1.69</td>
<td>183.32 1.24</td>
<td>181.80 0.98</td>
<td>156.00 1.21</td>
<td>155.49 1.09</td>
</tr>
<tr>
<td>Max</td>
<td>251.98 2.15</td>
<td>251.98 2.15</td>
<td>251.32 1.67</td>
<td>239.66 1.38</td>
<td>239.66 1.38</td>
<td>239.66 1.38</td>
</tr>
<tr>
<td>Mean</td>
<td>177.88 1.90</td>
<td>177.88 1.90</td>
<td>209.00 1.48</td>
<td>208.59 1.18</td>
<td>208.59 1.18</td>
<td>208.59 1.18</td>
</tr>
<tr>
<td>CV</td>
<td>0.16 0.06</td>
<td>0.08 0.10</td>
<td>0.07 0.11</td>
<td>0.04 0.04</td>
<td>0.04 0.04</td>
<td>0.04 0.04</td>
</tr>
<tr>
<td>r</td>
<td>0.21 -0.06</td>
<td>0.27 0.27</td>
<td>0.20 0.20</td>
<td>0.73 0.67</td>
<td>0.13 0.07</td>
<td>0.13 0.07</td>
</tr>
<tr>
<td>R²</td>
<td>0.04 0.00</td>
<td>0.07 0.07</td>
<td>0.04 0.04</td>
<td>0.53 0.45</td>
<td>0.45 0.45</td>
<td>0.45 0.45</td>
</tr>
</tbody>
</table>

Fig. 5-9. Plots showing the diurnal cycle of discharge and turbidity in comparison (left) and the correlation between both variables (right).
5.5 Hillshading during the ablation season

Figure 5-10 illustrates the development of shaded area on Rabots glaciär from May until September during peak temperature. A summary table including the solar position and calculated proportion of the shaded area is provided in table 5-7.

Fig 5-10. Maps showing the shadow extent at time of maximum temperature from May until September. The solar position was computed at the 15th of each month and the daily temperature maximum from the diurnal cycle of each month. The maps are in the coordinate system SWEREF99 TM. Two different DEMs were used for deriving the glacier contour lines (DEM 2011 by Keith Brugger) and the shadow extent (DEM 2008 by Lantmäteriet), whereas the glacier outline was mapped on the basis of a Landsat 8 image (acquired 2013-08-31).
The variability throughout the season is apparent with decreasing shadow extent starting from May, reaching the minimum during July (9.4% of the glacier is covered by shadow) and starting to increase gradually from then. The glacier dye tracing injection points are included in the maps. It is found that they are usually located outside the shaded area at time of peak temperature, except during September where experiments A6 and A7 are situated inside. The development of the shaded area and implications for the hydrology will be discussed in more detail later on.

Table 5-7. Results of hillshading analyses May until September. From the diurnal temperature cycle (for each month), the peak temperature $T_{\text{peak}}$ and time $t_{\text{peak}}$ was retrieved. The table further lists the azimuth and elevation at the 15th of each month that were used during the analyses. The last column gives the percentage of glacier area covered by shadow.

<table>
<thead>
<tr>
<th>Date</th>
<th>$T_{\text{peak}}$ [°C]</th>
<th>$t_{\text{peak}}$</th>
<th>Azimuth [°]</th>
<th>Elevation [°]</th>
<th>Shaded area [%]</th>
</tr>
</thead>
<tbody>
<tr>
<td>15 May</td>
<td>2.63</td>
<td>12:00</td>
<td>177</td>
<td>41</td>
<td>22.0</td>
</tr>
<tr>
<td>15 Jun.</td>
<td>4.61</td>
<td>13:00</td>
<td>195</td>
<td>45</td>
<td>12.3</td>
</tr>
<tr>
<td>15 Jul.</td>
<td>5.51</td>
<td>16:00</td>
<td>229</td>
<td>37</td>
<td>9.4</td>
</tr>
<tr>
<td>15 Aug.</td>
<td>4.82</td>
<td>15:00</td>
<td>227</td>
<td>30</td>
<td>16.9</td>
</tr>
<tr>
<td>15 Sep.</td>
<td>4.86</td>
<td>16:00</td>
<td>226</td>
<td>19</td>
<td>34.7</td>
</tr>
</tbody>
</table>
6 DISCUSSION

6.1 The Drainage System of Rabots glaciär

In this section the results of the dye tracing experiments in the proglacial stream and on the glacier, discharge measurements and catchment area analyses are going to be used to reveal characteristics of the drainage system. This includes the likely morphology of the drainage system as well as its temporal evolution. Along the section, the relation between different parameters (e.g. discharge and dispersivity) will be discussed.

6.1.1 Morphology

Based on the shape of the dye return curves and the relation between mean throughflow velocity ($v$) and dispersivity ($d$), the dye return data is categorized in different types of behavior. There is no good correlation between $v$ and $d$ for the glacier dye tracing experiments as supported by the low values for the coefficient of determination ($R^2=0.19$). Therefore, the experiments are examined individually.

There are different approaches in the literature to classify mean transit velocity and dispersivity. Theakstone and Knudsen (1981) established a threshold of 0.2 m s$^{-1}$ to distinguish between fast and slow flow. This value was also reported by Nienow (2011) by comparing various case studies. High flow velocities are indicative for channelized transport while low velocities suggest distributed water routing (Seaberg et al. 1988; Willis et al. 1990; Cotwon et al. 2013). Willis et al. (2009) regarded velocities of $<0.05$ m s$^{-1}$ as low, $\sim 0.1$ m s$^{-1}$ as moderate and $>0.15$ m s$^{-1}$ as high. Values for dispersivity are categorized in a similar way from $\sim 5$ m being low, $\sim 10$ m moderate and $>20$ m high. While low dispersivities indicate efficient water routing, high values for dispersivity are associated with less efficient drainage, possibly due to distributed drainage (Bingham et al. 2005; Nienow et al. 1998). Water that is stored either supraglacially or englacially may also increase the dispersion rates (Fountain 1993; Schuler et al. 2004).

Table 6-1 compiles values for mean throughflow velocity and dispersivity of a number of case studies in comparison to the results of Rabots glaciär. The lowest velocity was reported from Midtdalsbreen, Norway by Willis et al. (1990) with 0.007 m s$^{-1}$, and the highest was recorded on Haut Glacier d’Arolla by Nienow (1993) with 1.83 m s$^{-1}$. Values for dispersivity range between
DISCUSSION

0.044 m as noted on Storglaciären in the thesis by Ekblom Johansson (2013) up to 470 m on Dokriani Glacier (Hasnain et al. 2001).

In this study the injection points are grouped in three distinct categories as illustrated in Fig. 6-1. Type (1) includes experiments A3 and A7, which both exhibit values <20 m for dispersivity and <0.15 m s\(^{-1}\) for mean transit speed. Both dye return curves show a noisy pattern with several short-lived peaks, A7 being broader than A3. The values suggest that water routing occurred through a hydraulically relatively efficient but distributed drainage system following the discussion of Willis (2009). However, solely the form of the curve would suggest relatively channelized transport with a delay in release of meltwater that has been stored in the system. The broader peak width during experiment A7 compared to A3 is explained by the longer travel distance and supraglacial injection. Values for storage retardation are relatively small for both experiments (~5% for A7 and ~11% for A3). These low values may be due to predominantly supraglacial (A7) and subsequent englacial water routing. The higher storage retardation value for

### Table 6-1: Comparison of dye tracing results (throughflow velocity \(v\) and dispersivity \(d\)) from various glaciers. All injection locations were surface features (moulins, crevasses or supraglacial streams). Note that three different studies from Storglaciären, Sweden are included in the table.

<table>
<thead>
<tr>
<th>Site</th>
<th>Study</th>
<th>Number of successful tests</th>
<th>(v) [m s(^{-1})]</th>
<th>(d) [m]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rabots glaciär, Sweden</td>
<td>This study</td>
<td>8</td>
<td>0.04-0.28</td>
<td>2.89-92.63</td>
</tr>
<tr>
<td></td>
<td>Seaberg et al. 1988</td>
<td>13</td>
<td>0.03-0.16</td>
<td>1.5-82</td>
</tr>
<tr>
<td>Storglaciären, Sweden</td>
<td>Hock and Hooke 1993</td>
<td>10</td>
<td>0.07-0.29</td>
<td>2.3-55</td>
</tr>
<tr>
<td></td>
<td>Thesis by Fanny Ekblom Johansson (2013, not published)</td>
<td>25</td>
<td>0.03-0.65</td>
<td>0.044-26.67</td>
</tr>
<tr>
<td>Middalsbreen, Norway</td>
<td>Willis et al. 1990</td>
<td>15</td>
<td>0.007-0.228</td>
<td>0.7-41.6</td>
</tr>
<tr>
<td>Midre Lovénbreen, Svalbard</td>
<td>Irvine-Fynn et al. 2005</td>
<td>n.a.</td>
<td>0.08-0.57</td>
<td>0.07-22</td>
</tr>
<tr>
<td>Erikbreen, Svalbard</td>
<td>Vatne et al. 1995</td>
<td>9</td>
<td>0.07-0.32</td>
<td>2.9-41.8</td>
</tr>
<tr>
<td>Mittivakkat Glacier, Greenland</td>
<td>Memild 2006</td>
<td>8</td>
<td>0.10-0.27</td>
<td>5.8-19.6</td>
</tr>
<tr>
<td>Haut Glacier d'Arolla, Switzerland</td>
<td>Nionow 1993</td>
<td>57</td>
<td>0.03-1.83</td>
<td>n.a.</td>
</tr>
<tr>
<td>Unterarargletscher, Switzerland</td>
<td>Schuler et al. 2004</td>
<td>23</td>
<td>0.15-0.75</td>
<td>1.5-16</td>
</tr>
<tr>
<td>South Cascade Glacier, USA</td>
<td>Fountain 1993</td>
<td>36</td>
<td>0.01-0.32</td>
<td>n.a.</td>
</tr>
<tr>
<td>Brewster Glacier, New Zealand</td>
<td>Willis et al. 2009</td>
<td>8</td>
<td>0.04-0.16</td>
<td>2.3-28</td>
</tr>
<tr>
<td>Dokriani Glacier, India</td>
<td>Hasnain et al. 2001</td>
<td>10</td>
<td>0.10-0.47</td>
<td>25-470</td>
</tr>
</tbody>
</table>
A3 compared to A7 may indicate the use of another drainage pathway, either subglacially or even in the proglacial stream. If A3 and A7 are assumed to follow the same path, they are most likely disconnected from the subglacial system. This hypothesis is supported by the manual sampling results during A7, where dye was detected only in the southern stream (RJB). The low turbidity in this outlet stream supports the assumption of mainly supra- and englacial water routing. It may also be related to differences in sediment supply beneath the glacier, meaning that less sediment is available beneath the southern side of the glacier. However, this possibility is not likely, because no sediment differences were visible in the proglacial area.

Type (2), represented by experiments J1 and J2, has completely different characteristics with high values for v (0.28 m s\(^{-1}\) for J1; 0.21 m s\(^{-1}\) for J2) and d (86.24 m and 92.63 m respectively). The high velocities imply water routing through a channelized system, although the form of the curves indicates distributed water routing. The high dispersivity numbers suggest that the pathways are less hydraulically efficient. Perhaps this could be due to small channel size or roughness as pointed out by Willis et al. (2009). Dye retardation appears to be very high (71% and 59% respectively), which supports the previous assumption. Both dye return curves are characterized by remarkably long tails and pronounced noisy patterns, which would actually indicate distributed water routing. This contradiction is going to be taken up in the discussion later on. The long tail may be explained by the utilization of temporary storage zones such as back eddies, meander bends or regions with slowly flowing or even stagnant water as proposed.

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Fig 6-1. X-Y plots of dispersivity with mean throughflow velocity for the glacier dye tracing experiments.
by Willis et al. (1990). At times of low discharge dye may be trapped in storage spaces and eventually remobilized with increasing discharge. Unfortunately, no stream discharge was monitored during the July experiments. However, it is assumed to be rather high with large diurnal fluctuations in order to achieve the high throughflow velocities and temporary storage. Injection J2 was directly into a crevasse, from where it perhaps was transported englacially before reaching the subglacial system. In contrast, injection J1 was performed supraglacially, which is why the dye perhaps followed supraglacial channels before entering a moulin, being transported englacially and finally entering the subglacial system. The supraglacial transport might also significantly increase the overall velocity. It may also be possible that the return signal may not be attributed to the injection in the crevasse but from the supraglacial trace the day before (J1). This possibility and its consequences for the interpretation will be discussed in section 6.3.4.

Experiments grouped together as type (3) feature relatively high velocities (0.19 m s\(^{-1}\) - 0.25 m s\(^{-1}\)) and low values for dispersivity (4.25 m - 11.59 m). This configuration is indicative for channelized, hydraulically efficient water routing. The fast flow lowers the adhesion on sediments and thus the dispersivity as suggested for example Schuler et al. (2004). Since these experiments were performed within a time interval of two days, it is worth examining the three curves in detail. Fig. 6-2 shows the three breakthrough curves in the same plot with their respective time since injection. When just considering the shape of the curves, it is apparent that A1 is broadest (60.5 min), showing the least noise. Curve A2 is more narrow (30.5 min) and characterized by several, short-lived peaks. It is evident that A3 exhibits the same peaks, but more pronounced. Further, the curve appears to be even narrower (22 min). In accordance with Fountain (1993), multiple peaks indicate temporary storage or secondary pathways of dye before

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Fig. 6-2. Plot of experiments A1, A2 and A4.
returning to the main flow path. However, if they are small in size, they are not associated with carrying a high water volume. The observed pattern may be related to the varying discharge values, which decrease from 2.04 m$^3$ s$^{-1}$ in A1, 2.02 m$^3$ s$^{-1}$ in A2 to 1.57 m$^3$ s$^{-1}$ in A4. The computed storage retardation shows values of a similar magnitude A2 being the highest with 14%, followed by A4 and A1 each with 11%. The high discharge during experiment A1 resulted in the high throughflow velocity and comparatively large dispersivity values, perhaps because a larger surface was covered. In contrast A4 may reach the glacier snout at first due to the nearby injection point, the smallest curve width and transit time even though discharge was lower.

Experiment J3 is hard to categorize into the previous types, because dispersivity is higher than in groups (1) and (3) but still significantly lower than in group (2). The throughflow velocity of 0.15 m s$^{-1}$ is higher than in group (1) but lower than in the other groups. Storage retardation was calculated to be lower than for the experiments in group (2), but still remaining quite high (44%). The elongated tail is pronounced, not reaching base level throughout the duration of the experiment. This suggests that a considerable amount of dye is stored along its passage. With regard to Willis et al. (2009) this combination suggests channelized flow, being however hydraulically inefficient. However, compared to group (2), efficiency has increased relatively. The experiment was performed on 2013-07-22, thus in between the other experiments.

To summarize, there seem to be two distinct drainage systems on the northern and southern side of the glacier. On the northern side, the high throughflow velocities suggest channelized water routing while the curve shapes are typical for distributed flow. There is an increase in efficiency and channelization at lower altitudes and as the season progresses. This is going to be discussed more detailed in the following section. As the northern proglacial stream appears to be turbid it can be concluded that water is routed along the glacier bed. While the dye return curves from the southern side all look traditionally channelized, the numbers for A3 and A7 suggest water routing through a distributed but hydraulically efficient system. In contrast, experiments conducted at lower elevation on this side (A1 and A2) coincide with visual interpretations by showing efficient, channelized transport. There is reason to assume, that the dye follows englacial flowpathways inside the glacier without reaching the bed. Since the dye from experiment A7 was just picked up in the very clean outlet stream to the south, it seems apparent that the drainage system is disconnected from the subglacial one to the north. Reasons for this configuration may be found when considering then local topography, showing the extensive shaded part on the south of the
glacier at times of maximum melt throughout the season (Fig. 5-10, section 5.5). This could be a reason for less meltwater supply on the southern side compared to the northern one, which prohibits the development of a connection to the subglacial bed. Although not quantified, field observations indicate that the northern outlet stream seems to carry higher volumes of water compared to the southern one. This may be something to discuss in detail in a later study.

### 6.1.2 Seasonal development

A close analysis of seasonal development of the drainage system is rather hard, because none of the experiments were repeated at the same location due to the sub-optimal injection points used during July. When plotting dispersivity against the date of each experiment (Fig. 6-3) it turns out that there is a strong correlation between both ($R^2=0.89$). However, this result is biased since most of the experiments during the early season were conducted at higher altitudes, where higher dispersivities are expected.

The concept of increasing drainage efficiency had been reviewed in chapter 2.2.5 and is assumed to also occur on Rabots glaciär. As illustrated in the results, the diurnal cycles for each week showed two distinct peaks in week 30, which progressively occurred earlier in the day and evened out as the season progressed. The occurrence of two distinct peaks occurred with a time difference of 3 hours. This would imply that two pulses of discharge entered the proglacial system one after the other. A possible explanation would be the existence of different pathways that meltwater is following. The earlier pulse may originate from fast and efficient water routing; perhaps an already developed channelized system. Some of the meltwater probably took a different route, which delayed the discharge. This could be a flow through a distributed system or

![Fig. 6-3. Plot of dispersivity as the ablation season progresses.](image-url)
the temporary storage of water in the glacier. As the season progresses, the pattern of two distinct peaks becomes smoothed, meaning that there is no later release of meltwater. A range of incidents could account for that: for example an increase in hydraulic efficiency of the second pathway or a general change of the drainage pathways with rising discharge from week 30 to 31. The earlier timing of peakflow is clearly observable until week 36. Then, the diurnal pattern becomes blurred, eventually revealing baseflow conditions.

However, this examination of the hydrograph is rather subjective. Hannah et al. (1999; 2000) developed an approach to classify diurnal hydrographs with regard to their shape and magnitude by means of principal component and cluster analysis. The method attempts to differentiate between varying water inputs due to rainstorms and responses of the glacier system. It also allows evaluating the baseflow discharge. The discharge monitoring in the Rabots glaciär catchment did not start before the end of July, therefore the major part of the dynamics was not captured.

### 6.1.3 Catchment characterization

Runoff variability depends to a large degree on the percentage of the basin that is glacierized. As discussed during the literature review, catchments that exhibit a glacier cover of approximately 40% have the smallest variability in runoff. Rabots glaciär covers approximately 33% of the entire catchment. The estimated coefficient of variation is 0.38 for the monitoring season. It is logical that this value is significantly higher than the theoretical values suggested by case studies, because it is not representative for annual variations. Based on the study by Fountain and Tangborn (1985) the annual coefficient of variation with 33% glacier cover would be estimated to 0.1. Data from Switzerland and Austria showing a similar configuration was presented by Röthlisberger and Lang (1987). They concluded that basins with a glacier cover between 30 and 60% have a coefficient of variation <0.2. In any case, discharge monitoring of the entire hydrological year is required in order to draw conclusions about the runoff variability. This monitoring would be of great interest, because just a few catchments with this specific glacier proportion have been studied so far. Various studies found that year-to-year variations are lower for glacierized basins compared to basins without glacier cover (e.g. Braithwaite and Olesen 1988; Moore 1992). Due to the lack of data, no evaluations can be given concerning Rabots glaciär. However, it may be studied in the future.
6.1.4 Parameter dependencies

This chapter is going to discuss correlations between various parameters that were measured or computed during the dye tracing experiments. First, the relation between discharge and the meteorological parameters temperature and precipitation is examined. As seen in Fig. 6-4 the shapes of the diurnal temperature and discharge cycles are very similar; however exhibiting a time lag. The highest temperatures are recorded at 14:00 local time, whereas peak flow starts to occur 16:00, two hours later. This time lag of two hours characterizes the response time of the glacial system. Due to this time difference the coefficient of variation is just $R^2 = 0.49$, however, from visual interpretation the diurnal runoff cycle is forced by the daily temperature cycle (e.g. Willis 2005; Benn and Evans 2010)

The next plot (Fig. 6-5) includes runoff, temperature and precipitation for the entire monitoring season. Visual interpretations reveal that pronounced increases in discharge are related to precipitation events. The values for the coefficient of correlation are negligible, amounting $R^2 = 0.31$ for discharge and temperature and $R^2 = 0.01$ for discharge and precipitation. In order to distinguish between runoff from precipitation and runoff from ice melting, one would need further calculations. This includes the estimation of the aerial precipitation and also the total bulk discharge. The proportion of precipitation could then be subtracted in order to reveal the melt related runoff.

Fig. 6-4. Diurnal cycle of temperature (grey line) and discharge (black line)
The variation of the mean transit velocity with discharge is going to be investigated in the following.
With regard to Leopold and Maddock (1954), both hydraulic characteristics can be expressed as a simple power function. As seen in Fig. 6-6 throughflow velocity tends to increase with discharge both for the glacier ($R^2=0.65$) and river ($R^2=0.27$) dye tracing experiments. Velocities in the proglacial stream appear to be generally higher than in the glacial system. Table 6-2. compares the results with the ones from Storglaciären, Sweden by Seaberg et al. (1988) and from Midtdalsbreen by Willis et al. (1990). While Seaberg et al. (1988) also differentiated between proglacial and sub/englacial flow, Willis et al. (1990) divided the discharge recorded at the gauging station into proportions of 17 or 83%, attributed to the two tributary systems (T1 and T3). In the proglacial system of Rabots glaciär velocity increases as the 0.6 power of the discharge. This exponent is higher than the one from Seaberg et al. (1988). Leopold and Maddock estimated exponents of the south-western US and Great Plains to range between 0.07 and 0.55 with a mean of 0.34.

Table 6-2. Results from the velocity-discharge relationship analyses from Rabots glaciär, Sweden (this study), Storglaciären, Sweden (Seaberg et al. 1988) and Midtdalsbreen, Norway (Willis et al. 1990). The table compares multiplier, exponent and coefficient of determination of the power functions.

<table>
<thead>
<tr>
<th></th>
<th>multiplier</th>
<th>exponent</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>This study</td>
<td>proglacial</td>
<td>0.25</td>
<td>0.60</td>
</tr>
<tr>
<td></td>
<td>subglacial</td>
<td>0.02</td>
<td>3.44</td>
</tr>
<tr>
<td>Seaberg et al. 1988</td>
<td>proglacial</td>
<td>0.69</td>
<td>0.27</td>
</tr>
<tr>
<td></td>
<td>subglacial</td>
<td>0.26</td>
<td>1.00</td>
</tr>
<tr>
<td>Willis et al. 1990</td>
<td>T1 (83%)</td>
<td>0.06</td>
<td>1.00</td>
</tr>
<tr>
<td></td>
<td>T3 (17%)</td>
<td>0.04</td>
<td>0.60</td>
</tr>
</tbody>
</table>
Also the exponent of the (sub)glacial drainage appears to be significantly higher than the ones found in the other studies. However, the multiplier is on the same scale as the Midtdalsbreen T3 results. For a given discharge, the velocity of water in Rabots proglacial stream is two times less than the water in T3 draining Midtdalsbreen. According to Willis et al. (1990), this may be explained by a greater cross-sectional area and sinuosity of the Rabots glaciär drainage pathways. A more detailed investigation of the geometric and roughness characteristics of the pathways and their variation through the system goes beyond the scope of this study, but may certainly be interesting to look at.

The relation between elevation and throughflow velocity, dispersivity respectively is investigated in this paragraph. As visible in Fig. 6-7 there is a weak negative correlation between velocity and elevation, especially experiment A3 sticks out. Although located at a lower altitude, the velocity is quite low compared to the other experiments. As discussed above, this is related to the complexity of the most likely englacial, rather distributed drainage system. Moreover, the experiments were conducted at different times during the season, which is also assumed to affect the result. As suggested by many previous studies, there is an increase in drainage efficiency and transit time with time (e.g. Bingham et al. 2006; Copland et al. 2003; Cowton et al. 2013).

Fig. 6-6. Relation between transit velocity and discharge for the glacier and river dye tracing experiments. The experiments conducted July are not included in the plots, because discharge monitoring did not start before the end of July.
The other plot shows the correlation between dispersivity and elevation, with $R^2=0.35$. When examining the respective experiments it turns out that, even with a slight correlation between both parameters, the dynamics need to be explained by including the morphology of the glacial system and the timing of the experiments. As a conclusion, there tends to be an increase of dispersivity with elevation, but it will look different if the experiments are conducted early in the season or if the dye is transported mainly subglacially.

The parameters *storage retardation and dye recovery* are illustrated in Fig. 6-8. As shown, there is no obvious correlation between those two. The coefficient of determination amounts $R^2=0.3$ for the glacier dye tracing experiments with a positive correlation and $R^2=0.28$ for the river dye tracing experiments showing a negative relationship.

A general assumption could be that dye that was not recovered during the experiment is lost in the system, e.g. by temporary storage. If we expect all dye that was not recovered to be stored, this number would display in the storage retardation parameter. However, it appears that during a number of experiments, the weak dye recovery cannot be explained by temporary storage. This is particularly obvious in experiments A2, A4 and A7. During the first two experiments, the dye Uranine 33.3% was used, which is known to decay photochemically when exposed to daylight. For experiment A7 it seems odd to have low values for storage retardation and a very small dye
recovery at the same time. When examining the external conditions in detail, a possible explanation for the recovery rate could be refreezing of dye onto the ice. The experiment duration was the longest of all experiments (approximately 24h), thus including the night with falling temperatures and very low discharges. As discussed above, the dye was most likely following supraglacial and finally englacial routes, which supports the assumption that the loss of dye occurred through refreezing.

When examining the river dye tracing experiments, it appears that the proportion of dye that is computed to remain in temporary storage exceeds the dye that could not been recovered. This is the case for experiments D2, D3 and D5. A possible explanation could be poor mixing, which is, however unlikely, since at least two braided systems were included in each of the experiments. It may be more likely that the estimation for storage retardation based on the advection-dispersion model is not as accurate. It was adjusted to the best fit of the model efficiency criteria. Since just the falling limb is considered for computing the storage retardation, it may be overestimated. More investigations are needed regarding the relation between storage retardation and dye recovery, to potentially correct for the loss of dye due to photochemical decay.

As this section has shown, the glacier drainage system is very complex, which requires the careful consideration of the complete range of parameters.
6.2 Implications for the fate of hydrocarbon pollution

Assuming that soluble components of the aircraft fuel are transported with the glacial meltwater, some conclusions regarding pathways for pollution dispersal can be drawn from the dye tracing experiments. Dye in experiment A7 was injected supraglacially closest to the spill zone. It was detected only in the southernmost, less turbid proglacial stream and reached the peak concentration after approximately 15 hours in a single, well-defined peak. Storage retardation is estimated to be rather small on this side of the glacier, meaning that zones of immobile storage such as stagnant or slow moving water, back eddies or cavities (Willis et al. 1990) are missing. This suggests a relatively rapid transport through the southern system, which appears to be characterized by predominantly supraglacial and englacial flow pathways.

An automatic water sampler that was installed downstream nearby the gauging station collected water samples twice a day, which were analyzed in the laboratory to detect pollution traces. Traces were detected on only a few occasions, despite being collected every day. However, due to the probable storage and release mechanism operating downstream of the spill site, pollution is unlikely be detected by sampling at once-daily intervals. When it has been detected, we theorize that this coincided with a meltwater release event, or when exceptional melting or precipitation drives a flushing-out of the hydrological system.

From the relative clarity of the southern proglacial stream, pollution travelling through the glacier on the southern side of the drainage divide may not experience any substantial interaction with subglacial sediments and may therefore not be absorbed. This would then mean that pollution flushes through the glacier and reaches the proglacial stream without being stored. Sediments in the proglacial area could however store pollution on its emergence at the glacier front, particularly due to the braided nature of the proglacial system. Absorption into subglacial sediments may be more important on the northern side of the glacier if fractions of the hydrocarbons are evaporated and precipitated on this side of the glacier, or travel there through supraglacial flow.
6.3 **Uncertainties and Limitations**

The data and methods that were used during the thesis may contain some errors or limitations, which can have impact on the quality of the results. They are going to be examined and discussed in this chapter.

6.3.1 **Experiment set up**

It is somewhat hard to draw conclusions about the drainage system of Rabots glaciär since just a little part of the season was captured. Due to the early onset of melt, a major part of drainage system evolution from inefficient to efficient most likely occurred before the monitoring started. By just examining the hydrograph between the end of July and the beginning of September, some of the important dynamics were most likely not reported.

None of the injection points was used twice as the July injection points were not optimally chosen, which makes it impossible to infer changes of the drainage system originating from the same point. Further, no discharge estimations were available for the July experiments. Thus, some of the parameters (e.g. dye recovery) could not be calculated for them.

6.3.2 **DEM & Mapping Accuracy**

According to USGS ([http://landsat.usgs.gov/landsat8.php](http://landsat.usgs.gov/landsat8.php), 13 Jan., 14), data of the Operational Land Imager (OLI) sensor, which is carried by the Landsat 8 satellite, exhibits a 12 m circular error with 90% global confidence accuracy. The data is delivered with 15 m to 30 m pixel sizes depending on the respective bands. After pansharpening the image, mapping resolution increased to 15 m. The different band combinations allowed a good differentiation between snow and ice as well as the other land cover types. One should bear this resolution in mind when working with the glacier extent data. In order to compare the extent to the following year, the data may not be accurate enough. However, for the purpose of this thesis, i.e. producing an overview map, it is sufficient.

The various DEMs were used for different purposes. The DEM provided by the Swedish mapping agency Lantmäteriet was created in 2008 and has a 10 m resolution. From this time, glacier changes are most likely to be significant, which is why the DEM was solely used for the hillshade analysis. The DEM provided by Keith Brugger from 2011 has also a 10 m resolution.
and seems to agree with the current glacier extent much better than the previous one. It was used for producing the contour lines for the basemap. In order to perform catchment analyses, a DEM was needed that covered the entire catchment area. The ASTER GDEM, released in 2011, meets this requirement. As mentioned earlier the data has an error of 17 m at 90% confidence interval and may exhibit undulation effects in the vertical dimension. The pixel size turned out to be 20 m. There may be errors in the watershed analysis associated with this quite coarse resolution and the probability of errors. However, to get an idea of the approximate catchment area and the proportion that is covered by Rabots glaciär, this DEM seems to be sufficient.

6.3.3 Disturbances of the dye tracing signal

Turbidity is known to influence the accurate measurements of dye concentrations. This was particularly noticeable when analyzing the manual samples from the northern and southern outlet streams of Rabots glaciär. The northern stream, which appears to be very turbid, showed large fluctuations in the dye return signal between 3.78 ppb and 15.1 ppb without a clear trend. This turbidity noise could be misinterpreted as actual signal. Since it is following a diurnal cycle (see results section 5.2.3), it is subject to change constantly, which makes it hard to define the noise level. Schnegg (2003) points out that the particles in the water scatter the incoming solar radiation in all directions, also towards the detectors of the FL30 instrument. Therefore, it is important to use enough dye to show up in the signal clearly. The existence of organic matter may also influence turbidity and thus the signal scattering. However, it is not further discussed in this context, because no significant amounts are assumed to occur in the streams.

Temperature is also known to have an impact on dye, with a temperature increase of 20°C lowering the dye concentration by approximately 25% (Turner Designs 2013, User’s Manual). Therefore, it is required that samples and calibration solutions adjust to the same temperature before starting the analyses. Since this fact was considered during the laboratory work, potential impacts can be ruled out.

6.3.4 Misinterpretations of the FL30 signal

It is possible that errors occur due to misinterpretations of the recorded signal. Dye may become trapped in the glacier system and be released later on (Seaberg et al. 1988; Willis et al. 1990; Fountain 1993). This may include freezing onto the ice supra- or englacially but also temporary storage in the subglacial system or adhesion to sediments. The trapped dye may be released
through melting or increasing discharge at a later time. If several experiments are conducted successively, the delay of dye may influence the results of other experiments, for example by superimposition.

The interpretation of the dye return curve is subjective, which is why errors cannot be ruled out completely. For most of the experiments, the time interval between several experiments was quite long to assure full dye recovery. A common approach is to perform experiments at lower altitudes first before moving upwards, because transit times and complexity of the drainage system are expected to increase with elevation. Moreover, different dyes (Uranine 33.3% and RWT 20%) were used simultaneously to avoid interferences of the signals.

However, for the first experiments in July (J1 and J2) the possibility of misinterpretations cannot certainly be excluded. Figure 6-9 illustrates the signals that were recorded between 2013-07-04 and 2013-07-05. There is no data available during the night between both days. However, as data interpretation shows, the signal had not reached the baseline concentration when the FL30 instrument was turned off. Therefore, it is likely that dye signal did still pass through the gauging station without being recorded. Based on the parameter calculations, the storage retardation is estimated to be approximately 71%. The peak was recorded 16:14 (local time). Although no discharge data is available, it is reasonable to assume that discharge decreased some time during late afternoon and remained low during the night. Since the dye was injected supraglacially, it probably followed a number of pathways before entering the subglacial system at lower elevations. It is feasible that some dye got trapped in the ice and was remobilized by melting the

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![Fig. 6-9. Plot of the FL30 signal that was recorded between 4th and 5th July 2013. The arrows mark the times of injection.](image)
next day. As discussed by Fountain (1993) dye may also be passed to temporary storage at low discharges and be mobilized when runoff starts to rise again. Since the signal, which was attributed to experiment J2, was recorded in the afternoon, a time where discharge is known to rise, the signal may belong to experiment J1 or superimpose the signal of experiment J2. The fact that the recorded concentrations are very low (1.43 ppb as peak concentration) supports this assumption.

Supposing that the FL30 instrument did not pick up any signal of the J2 injection, the calculations would be incorrect. The mean transit velocity would by significantly lower, i.e. 0.06 m s\(^{-1}\) or less (assuming that dye emerged after the monitoring had finished). The water routing would not occur through channels as previously assumed, but through rather distributed pathways, perhaps with low hydraulic efficiency. Further, it may not be likely to find channelized flow at this high altitude in early July.

### 6.3.5 Estimation of dye recovery

Dye recovery could only be computed for the August experiments, because no discharge data was available in July. For the river dye tracing experiments dye recovery amounts to between 55.05% and 93.52% with a mean of 79.39%. The glacier dye tracing experiments (August) show recovery rates from 8.25% to 78.69% with a mean of 34.82%. This low average is mainly influenced by the poor recovery rates of the Uranine 33.3% experiments. This dye is prone to decay when exposed to daylight (e.g. Leibundgut et al. 2009). Other sources of dye loss are temporary storage in the subglacial system, adhesion on sediment particles or freezing onto ice. These different processes were discussed when interpreting the return curves and drawing conclusions about drainage system morphology.

### 6.3.6 Sampling rate

Nienow et al. (1996a) investigated the effects of the sampling interval on the breakthrough curve properties, such as throughflow velocity, dispersion coefficient and dispersivity. They found an increase of the dispersion coefficient and the related dispersivity with higher sampling intervals, while no systematic difference was observed for the throughflow velocity. They determined that the sampling rate should be lower than 1/16 of the residence time (i.e. duration between dye injection and peak concentration) in order to obtain reliable results.
As noted in table 6-3 the minimum required sampling interval was met during all the experiments. The FL30 instrument recorded a signal every 30 s, whereas for all experiments a sampling interval of at least 2 min would be acceptable. The manual sampling of experiment A7 took place every 20 min in the beginning, followed by 10 min sampling intervals once dye was recorded at the gauging station. Those time intervals still remain above the ideal sampling rate and thus reliable estimates of the dye return curve properties are possible.

6.3.7 Manual vs. automated sampling

In Fig. 6-10 the results from the manual sampling in stream RJB are plotted against the recordings from the FL30 instrument. Both datasets show a very good correlation with $R^2=0.85$. Because there are significant differences in the range of values, two vertical axes were needed. The curve of the manual sampling has a significantly higher range up to 10.78 ppb compared the
range of the FL30 recordings (peak of 2.54 ppb). It was hard to define a noise level due to the large fluctuations. This is why values partly exhibit negative values. Furthermore, the curve has a noisier pattern compared to the return curve from the automatic sampling, perhaps related to the instrument AquaFluor being very sensitive to scratches on the cuvettes. Also the results from the dye tracing calculations differ regarding the estimated dye recovery, dispersion coefficient, dispersivity and storage retardation respectively. All values tend to be higher for the manual sampling. As already indicated, this may be related to turbidity, which influences the signal. The FL30 instrument accounts for it automatically whereas a correction is missing for the analyses with the handheld AquaFluor.

6.3.8 Dye tracing calculations

There are some weaknesses of the dye tracing calculations. First of all, it is apparent that a straight line as transit distance underestimates the actual distance, because actual pathways may be sinuous. Related to this, the mean throughflow velocity is also a minimum estimate. The advection-dispersion model, which was used for computing the dispersion coefficient, showed a satisfactory performance for most of the experiments. Exceptions are experiments J1 and J2, where the model did not capture the dynamics. To solve this issue, another model that accounts for storage-retardation processes could be applied. Schuler (2002) used a mobile-immobile model that accounts for heterogeneous flow conditions.

Furthermore, the calculated values often suggest different drainage configurations regarding efficient/channelized and inefficient/distributed flow than expected from visual interpretations. Another issue encountered is that supraglacial transport increases the overall velocity and therefore influences the results of throughflow velocity and dispersivity. As discussed, the dye return curve of experiment J1 looks typically distributed whereas the high transit velocity suggests channelized flow. The high numbers of dispersivity are typical for hydraulically inefficient transport, which is found in distributed systems. Due to the supraglacial injection, the transit velocity may have been biased and flow may actually be distributed. The other experiment that shows a contradiction is experiment A3 showing a breakthrough curve typical for efficient and channelized flow. Whereas the low dispersivity supports the assumption of efficient transport, the very low transit velocity suggests distributed water routing. A possible explanation could be delay due to storage mechanisms in the englacial system and subsequent sudden release of
meltwater. This may also be the case for experiment A7, where the curve form does not agree with the throughflow velocity.

6.3.9 Rating curve accuracy

In order to create rating curves from relative gauging height and water pressure, just five discharge measurements were available, all of them conducted in August. This is a potential error source, because they may not reflect flow conditions of the entire ablation season. As it turned out, the rating curve had to be extrapolated in order to compute the hydrographs. As highlighted in several studies (e.g. Fenton and Keller 2001; Fenton 2001), channel geometry plays a significant role, especially for estimating high flow. However, this has not been taken into account in this study, because none of the parameters were measured in the field. Generally, the accuracy can be improved by using as many direct measurement points as possible (Rantz et al. 1982).

Figure 6-11 compares the hydrographs derived from relative stage and from water pressure. The agreement between both curves is evaluated by means of different criteria (table 6-4). Both curve agree very well as proved by visual interpretations and the statistics. However, there are some deviations where the discharge derived from gauging height appears to exceed the hydrograph.

Table 6-4. Criteria for the agreement between discharge derived from relative gauging height and from water pressure.

<table>
<thead>
<tr>
<th></th>
<th>R²</th>
<th>E</th>
<th>ICA</th>
<th>PBIAS [%]</th>
<th>RMSE</th>
<th>RSR</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0.96</td>
<td>0.95</td>
<td>0.99</td>
<td>4.71</td>
<td>7.03</td>
<td>0.23</td>
</tr>
</tbody>
</table>

Fig. 6-11. Comparison of discharge derived from relative gauging height and from relative water pressure.
from water pressure. There seem to be some systematic deviations in the beginning, around the middle of August and towards the end of August, which are rather small. However, some deviations stand out, on 2013-08-23 and some days from 2013-08-29 in particular. In these cases, the estimations from gauging height are some orders of magnitude higher than the ones derived from water pressure. None of the events are associated with high precipitation events. Therefore, a possible explanation for this could be temporary errors or inaccuracies within the instruments. Another, more likely reason could be a change in channel geometry that led to variations in water height or flow rate.

6.4 Future Research

The discussion gave rise to some future research directions, which are going to be collected and discussed in this section.

1. In order to rule out the uncertainty regarding experiments J1 and J2, it would be recommended to perform similar experiments again in the following season. The dye tracing experiments already followed the longitudinal profile, however there was a significant time lag between the experiments on the northern side of the glacier. In order to explore the morphology and drainage connectivity with elevation, it would be beneficial to follow a profile approximately at the same time in the season.

2. More investigations are needed with regard to the velocity-discharge relationship to reveal characteristics about the hydraulic geometry of the drainage pathways. Repeated dye tracing experiments following the diurnal cycle as done by Nienow et al. (1996b) or Schuler (2002) are needed. Further, the discharge entering the moulin should also be monitored and quantified.

3. The peak dye concentrations picked up by the FL30 instrument ranged between 1.43 ppb and 52.89 ppb. Generally, the amount of dye should be increased to avoid approaching the lower detection limit of the instrument. The earlier in the season or the higher the elevation, the more dye is needed. Following the experiments during 2013 we now know more about the necessary injection volumes for optimal dye return concentrations in the future.
4. Locating the FL30 instrument near the gauging station, as done in August, was quite practical. It was possible to correct for transit distance and travel time in the stream. However, if pursuing this practice, the use of Uranine 33.3% should be reconsidered, because it is prone to photochemical decay while traveling in the stream.

5. Two drainage systems have been identified, which are perhaps distinct from each other. Manual sampling in both streams would be recommended during more experiments to verify this hypothesis. Further, the estimation of the volume draining from each outlet stream would be useful.

6. In order to enhance the accuracy of the rating curve it is recommended to increase the number of the discharge experiments in the proglacial stream. It is particularly important to capture low and high flow conditions.

7. The approach by Hannah et al. (1999; 2000) allows an objective evaluation of hydrograph shape and magnitude. It is possible to distinguish between inputs due to precipitation and melt related water input and thus reveal baseflow conditions. If possible it would be useful to cover the entire hydrological year or at least the complete ablation season. Discharge recordings of an entire year would further enable conclusions about discharge variability, which then can be assessed in relation to the percentage glacier cover of the catchment.

8. The monitoring program should be continued for capturing inter-annual variability. It may prove useful to use approximately the same injection locations to detect differences or changes in the drainage system.

9. An interdisciplinary research project involving glacier hydrology investigations and ground penetrating radar could answer questions regarding the existence of a cold ice surface layer and its implications for glacier hydrology. Moreover, in order to explore subglacial drainage morphology, it could be useful to include field studies of sedimentology in the proglacial area.

10. Investigations of how dye dissolved in water and dye dissolved in kerosene (preferably the same fuel the Hercules airplane was refuelled with) percolates through a snowpack provides information on how dye tracing experiments can be used as a proxy for pollution propagation through the glacier system. Analyses of ice cores could further improve our knowledge on transport and storage mechanisms of the hydrocarbon pollution.
Investigations of Rabots glaciär’s drainage system during the ablation season of 2013 were performed through river and glacier dye tracing experiments, monitoring of discharge in the proglacial stream and analyses of the catchment by using digital elevation models.

Analyses of the proglacial discharge showed an earlier timing of peak flow as the season progressed. A return to baseflow conditions most likely occurred after week 36. From numerical analysis of the dye returns, the hydrological configuration is characterized by an efficient drainage system with both channelized and distributed flow pathways at the end of the ablation season. The drainage efficiency and channelization increased as time passed and with decreasing elevation. Drainage is divided into a system dominated by subglacial water routing on the northern side and a system, which is proposed to be characterized by englacial flow on the southern side. Investigations of the glacial thermal profile may help to explain the proposed missing connection to the glacier bed on this side. There are indications that the two systems are disconnected from each other in the englacial and subglacial domain.

The results suggest that pollution that reaches the meltwater from the spill side is rapidly transported through the southern drainage system dominated by englacial flow pathways. The storage and release mechanisms encountered imply that it is unlikely to detect pollution traces in the water samples collected downstream unless they capture peak concentrations. The interaction with subglacial sediments is assumed to be rather low on this side of the glacier, however, absorption of pollution may occur in the proglacial area.

Some uncertainties and limitations have been identified regarding the interpretations of the dye return curves or the accuracy of the calculations. If recorded nearby the gauging station, Uranine 33.3% is not suitable for this environment due to photochemical decay in the proglacial stream. Further, the calculated values suggest different drainage configurations with regard to efficient/channelized and inefficient/distributed flow than expected from visual interpretations. Future investigations should focus on confirming and constraining the divergence of the drainage system, by sampling in both main outlet streams during the dye tracing experiments. Discharge monitoring of the entire season would be required to capture the evolution of the drainage system. An interdisciplinary approach including ground penetrating radar could reveal information regarding the link between thermal regime and hydrology.
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REFERENCES


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Glacier dye tracing experiments together with their respective injection location