HYDROECOLOGICAL RESPONSE OF ARCTIC RIVERS TO CLIMATE CHANGE

by

PHILLIP JOHN BLAEN

A thesis submitted to the
University of Birmingham
for the degree of
DOCTOR OF PHILOSOPHY

School of Geography, Earth and Environmental Sciences
College of Life and Environmental Sciences
University of Birmingham
March 2013

UNIVERSITY OF
BIRMINGHAM

University of Birmingham Research Archive

e-theses repository

Abstract

Although the Arctic is one of the most vulnerable regions to climate change, links between hydrology and ecology in high-latitude northern river basins are not well understood. Interdisciplinary research over three summer melt seasons (2010 to 2012) in the Kongsfjorden area of northwest Svalbard identified process connections between conceptual water sources, physicochemical habitat, and ecological structure and functioning in river basins. Water source dynamics determined from hydrochemical and isotopic data indicated differences in meltwater and groundwater contributions to river flow which varied both spatially and temporally at seasonal and year-to-year timescales. Non-glacier-fed rivers were characterised by less variable flow regimes, warmer water temperature, lower suspended sediment concentration and more stable channel morphologies. Several physicochemical habitat variables, notably water temperature and channel stability, were related significantly to rates of nutrient uptake and macroinvertebrate community structure. These data suggest that a future shift towards groundwater-dominated flow regimes may increase biotic diversity and rates of nutrient cycling in some high-latitude rivers. Key research findings are synthesised in conceptual models and provide a framework to understand the hydroecological response of these Arctic river systems to climate change.

Acknowledgements

This research was made possible by support from many people and organisations.

Firstly, I would like to thank my supervisors, Sandy Milner and David Hannah at the University of Birmingham, and Lee Brown at the University of Leeds, for their guidance and friendship during the past four years. I also thank the UK Natural Environment Research Council (studentship NE/G523963/1) and the Research Council of Norway (219105/E10) for supporting the project financially.

I am grateful to the Alfred Wegener Institute for Polar and Marine Research and to Fred Wenger at the Norwegian Water Resources and Energy Directorate for providing hydroclimatological data.

My thanks go to Gillian Kingston for her assistance with hydrochemistry analysis and to Ian Boomer for analysing the isotope samples. Rebecca Sutton was very generous in allowing me access to laboratory facilities at the University of York, as was Brigitte Lods-Crozet at the Musée de Zoologie in Lausanne for verifying the chironomid specimens. I am also particularly grateful to Richard Johnson for supplying the necessary field equipment and for his continual patience when I kept asking for more.

A large proportion of this project was spent in Svalbard, and as such I owe a great deal of thanks to all of the scientists in Ny Ålesund and staff of Kings Bay AS for their help during my time there. Nick Cox at the British Antarctic Survey deserves a special mention for providing both logistical support and an inexhaustible supply of polar-related stories that never failed to make me smile. My thanks also go to Aga Nowak-Zwierz, Fiona Moore and Tris Irvine-Fynn for their company and support in the field, to Steve Coulson at The University Centre in Svalbard for ensuring my water samples stayed frozen, and to John Brittain at the Natural History Museum in Oslo for his help with obtaining a grant which led to an extra fieldwork season.

There are many other people whose assistance and ideas have been invaluable, but in particular I need to thank Kieran and Grace for discussions over many cups of tea which improved this research immensely.

Finally, I am indebted to my family and Kerstin for their endless support, without which none of this would have been achieved.

Thank you all.

Table of contents

List of Illustrations

List of Tables

CHAPTER 1

INTRODUCTION, RATIONALE AND THESIS AIMS

1.1 Rationale and research gaps

The Arctic is one of the most vulnerable regions of the world to climate change (McBean *et al.*, 2005; Anisimov *et al.*, 2007), with accelerated warming documented in recent decades (Serreze *et al.*, 2000; Walsh *et al.*, 2011). Mean summer air temperature in the Arctic is now higher than at any point during the past 2000 years (Kaufman *et al.*, 2009; Walsh *et al.*, 2011). Climate models forecast continued warming of 5-8 ºC during the rest of the 21st century (Anisimov *et al.*, 2007; Vavrus *et al.*, 2012) coupled with increased annual precipitation (Kattsov *et al.*, 2005, Mahlstein *et al.*, 2012) and a rise in the frequency and magnitude of extreme climatic events (Christensen *et al.*, 2007; Brown *et al.*, 2008). Such changes are likely to have implications for sea ice thickness and coverage, snowpack volume, permafrost extent and glacier runoff in polar regions (Walsh *et al.*, 2005; Hinzman *et al.*, 2005; Comiso *et al.*, 2008; Callaghan *et al.*, 2011a) due to due to strong linkages and complex feedback interactions between atmospheric, cryospheric and hydrological systems (Serreze et al, 2009; Callaghan *et al.*, 2011b).

Arctic environments are potentially valuable indicators of global environmental change due to high sensitivity to climate variability. In this context, high-latitude river systems are sentinel systems because they integrate multiple basin-scale processes potentially modified by climate forcing (Milner *et al.*, 2009). Hydrological and temperature changes could alter nutrient and sediment budgets and thus modify physicochemical habitat conditions (*sensu* Hannah *et al.*, 2007). In turn, these could influence freshwater biota and affect rates of ecosystem functioning in Arctic rivers (Prowse *et al.*, 2006; Milner *et al.*, 2009; Woodward *et al.*, 2010). However, our understanding of the response of these systems to climate change is limited, despite a growing need for research to inform basin management decisions, assess freshwater, nutrient and sediment delivery to the Arctic Ocean, and ensure water security for northern communities (Peterson *et al.*, 2002; Barnett *et al.*, 2005; Evangard *et al.*, 2011; Holmes *et al.*, 2012).

This study focused on key aspects of Arctic river ecosystems potentially modified under a changing climate, and for which we lack detailed information at present. These were investigated to achieve a more comprehensive understanding of the hydroecological response of Arctic rivers to climate change. The following research gaps for investigation in this study were identified following an evaluation of the literature in Chapter 2:

1. High sensitivity and strong forcing between atmospheric, cryospheric and hydrological systems make high-latitude rivers highly vulnerable to climate change. Nevertheless, the hydrological processes and key water sources that govern river flow dynamics in Arctic river basins, and how these vary over seasonal and inter-annual timescales, are not well understood (Hodgkins *et al.*, 2009; Cooper *et al.*, 2011).
2. Climate warming is likely to modify water temperature regimes in high-latitude regions in the foreseeable future. However, thermal regimes of Arctic rivers are not well studied and research is required to characterise and assess the drivers of water temperature dynamics in these systems (Irons and Oswood, 1992; Lammers *et al.*, 2007; Chikita *et al.*, 2010).
3. Future changes in the nutrient dynamics of Arctic rivers are highly uncertain as increased nutrient loading associated with permafrost degradation may be mitigated by increased biotic activity and nutrient uptake associated with warmer water temperature. The processes controlling rates of nutrient uptake in these river systems require further study (Petrone *et al.*, 2006; Holmes *et al.*, 2008).
4. Shifts in water sources contributing to Arctic river flow regimes could modify physicochemical habitat conditions, particularly water temperature, which may in turn influence biotic community structure and function. There is now a need to quantify the linkages between these hydrological and ecological systems to gain a comprehensive understanding of the response of Arctic rivers to climate change (Prowse *et al.*, 2006; Wrona *et al.*, 2006; Vincent *et al.*, 2011).

1.2 Aim and objectives

In light of these research gaps, the overarching aim of this thesis is to adopt an interdisciplinary approach to understand process connections between water sourcing, aquatic habitat conditions and ecological structure and functioning in Arctic river basins, and thus assess the vulnerability of these systems to climate change.

To address each of the identified research gaps, the objectives of the study were to:

1. Identify the major hydrological processes and key water sources that govern river flow dynamics in Arctic river basins (Chapter 4).
2. Quantify, and assess the drivers of, water temperature dynamics in Arctic river systems (Chapter 5).
3. Calculate, and identify the processes that control, rates of nutrient uptake in Arctic river systems (Chapter 6).
4. Understand the processes that link basin water source contributions, physicochemical habitat conditions and biotic community structure in Arctic rivers (Chapter 7).

1.3 Thesis structure

The thesis layout is discussed below (Figure 1.1).

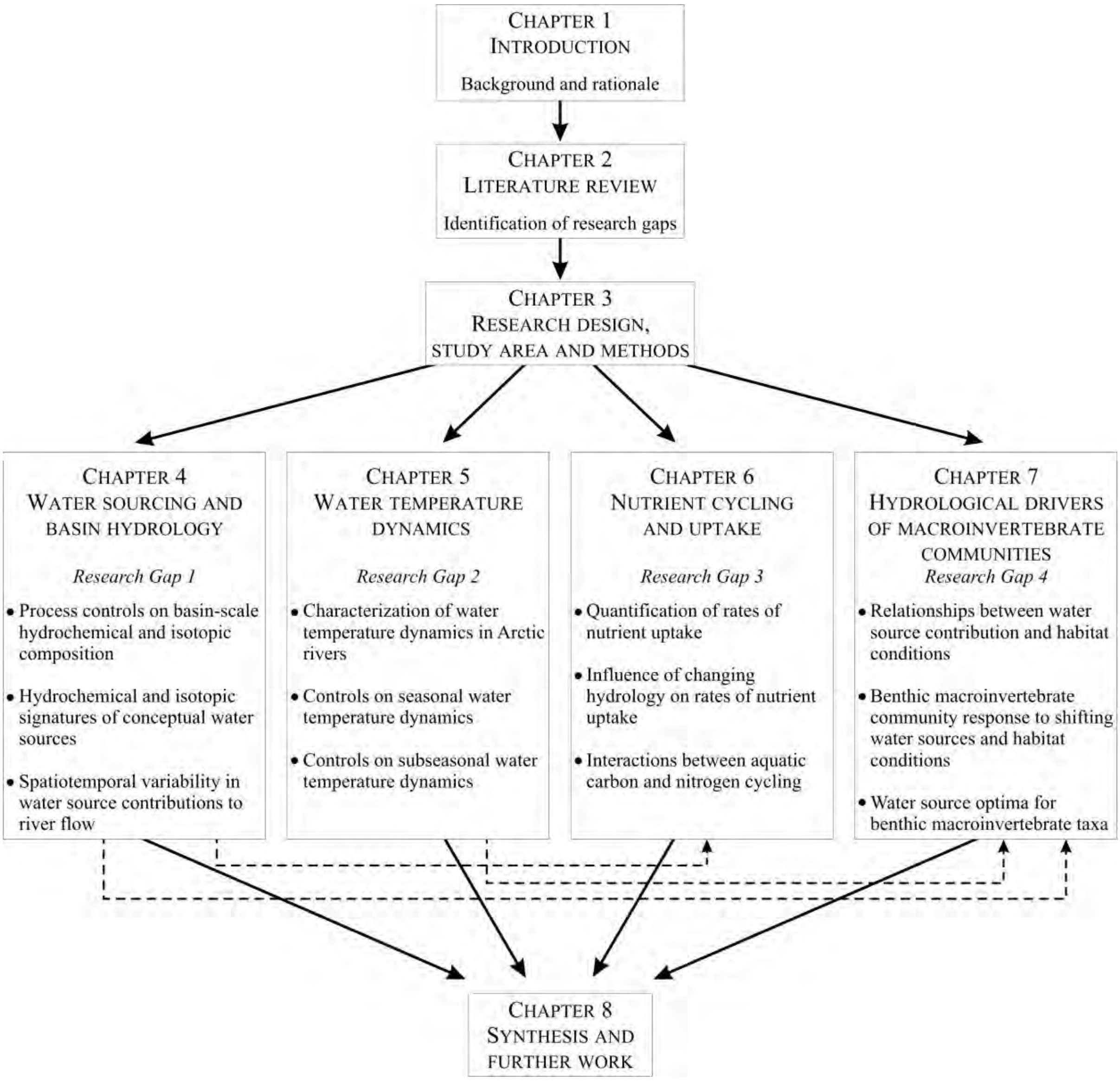


Figure 1.1: *Overview of the thesis structure. Arrows show inter-relationships between chapters.*

The literature review in Chapter 2 identifies the research gaps detailed in section 1.2. The research design and an overview of the data and sampling methods used in the thesis are presented in Chapter 3. Basin water sourcing and the hydrological processes that govern flow regimes in Arctic rivers are examined in Chapter 4, thus addressing the first identified research gap. Following this, water temperature dynamics across multiple Arctic river basins are explored in Chapter 5 where these data are used to infer the major drivers of thermal variability in high-latitude rivers, hence addressing the second identified research gap. Work in Chapter 6 draws on the results in Chapter 4 to address the third identified research gap by investigating the impact of shifting water sources on nutrient cycling within Arctic rivers. Similarly, the influence of hydrological change on physicochemical habitat conditions and benthic macroinvertebrate communities in Arctic rivers is presented in Chapter 7, thus addressing the final research gap. The thesis is concluded by Chapter 8 which synthesises the major findings of the research. The wider implications of the results are discussed in the context of ongoing and future environmental change in Arctic regions, and areas for further research are identified.

1.4 Chapter summary

A background and rationale to conduct research on the hydroecology of Arctic river basins has been provided. Key research gaps and objectives to address these have been documented, and an overview of the thesis structure has been outlined. A literature review which contextualises the thesis and identifies research gaps to be addressed in subsequent chapters is presented in Chapter 2.

Some research from this thesis has been peer-reviewed and published. A *Hydrological Processes* journal article was published from research in Chapter 5 (Appendix A1). Research in Chapters 4, 6 and 7 has been accepted for publication in *Hydrological Processes*, *River Research and Applications*, and *Freshwater Biology*, respectively. Research in Chapters 4 to 7 has also been presented at conferences and workshops.

CHAPTER 2

CLIMATE AND CRYOSPHERIC INFLUENCES ON ARCTIC HYDROECOLOGY

2.1 Introduction

A rationale for the research undertaken in the thesis was provided in Chapter 1 by highlighting the wider importance of Arctic river systems as indicators of environmental change in high-latitude regions. Here, a more detailed review of contemporary understanding of Arctic hydroecology is provided. The chapter begins with an overview of current and expected future climatic change relevant to Arctic river ecosystems, and continues with four separate sections which examine water sourcing and Arctic basin hydrology, river temperature dynamics, nutrient cycling and uptake, and drivers of aquatic macroinvertebrate community structure, respectively. The chapter concludes by summarising the identified research gaps and presents the major objectives of the thesis.

2.2 Environmental change in the Arctic

Arctic river systems are vulnerable to the effects of large-scale climate change due to complex and strongly synergistic feedbacks between climate, cryospheric and river ecosystems at high latitudes. Work in this section provides background information on recent observations and future changes in these different environmental systems to contextualise the aspects of high-latitude hydroecology considered later in this study.

2.2.1 Climate

The Arctic climate is characterised by high seasonality, with cold dark winters and long days during summer months (McBean *et al.*, 2005). Snow and ice with high albedo reflect large amounts of incoming short-wave radiation and maintain low air temperature at high latitudes (Walsh *et al.*, 2011).

Although a dynamic region historically (Figure 2.1; Kaufman *et al.*, 2009), Arctic surface air temperature has risen at around twice that of the global mean for several decades (McBean *et al.*, 2005; Screen and Simmonds, 2010). The rate of warming has accelerated in recent years and there has been an increase in the frequency of daily temperature extremes (Figure 2.2; Serreze *et al.*, 2000; Walsh *et al.*, 2011). Long-term variability in Arctic air temperature has been attributed largely to low-frequency (>10 y) changes in atmospheric circulation patterns, chiefly the Arctic Oscillation (AO) and Pacific Decadal Oscillation (PDO) (Overland *et al.*, 2008). However, the warmest years on instrumental records occurred in the 2005 to 2009 period when neither the AO nor PDO were in states conducive to high-latitude warming (Walsh *et al.*, 2011), suggesting that anthropogenic climate change is now having a discernable effect on the Arctic climate system (Screen and Simmonds 2010; Cohen *et al.*, 2012).

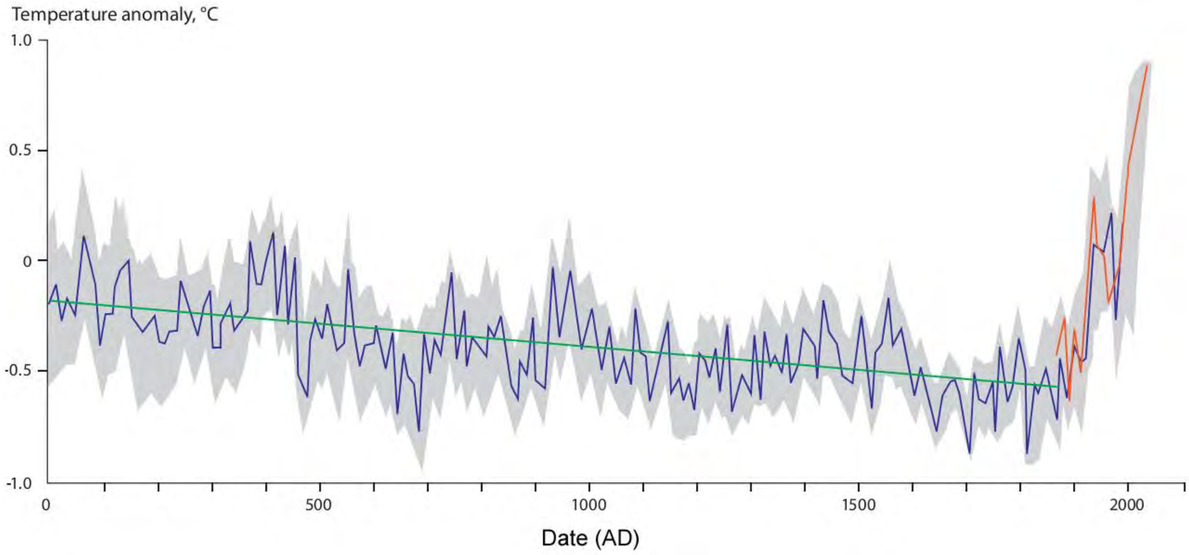


Figure 2.1: *Estimated Arctic average summer air temperature anomalies (ºC) for the past 2000 years relative to the 1961–1990 mean (blue line), based on proxy records from lake sediments, ice cores and tree rings at 23 sites. The grey area represents variability around the mean. The green line shows the long-term trend from 0 – 1900 AD. The red line shows the instrumental record.*

From Walsh et al. (2011)

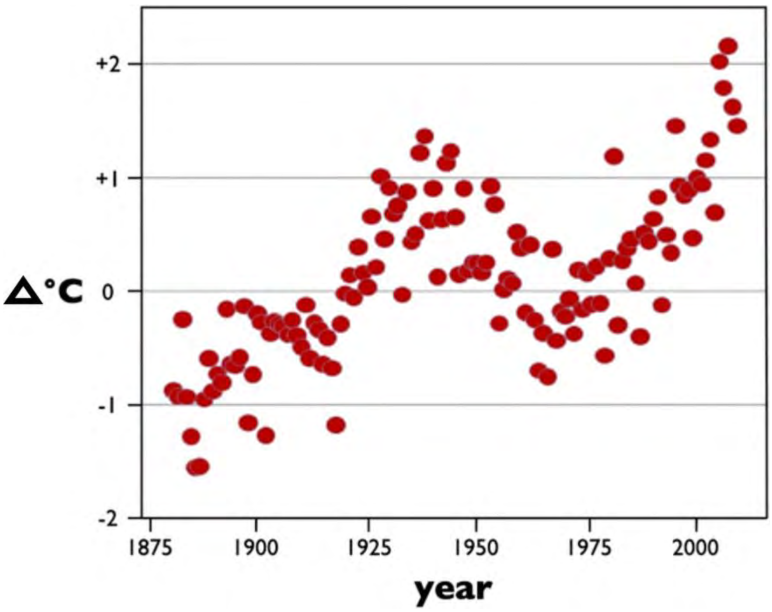


Figure 2.2: *Annual Arctic air temperature anomalies (ºC) for 1880–2009, averaged over 60º–90ºN, relative to the mean for 1961–1990. Based on station observations of surface air (2 m) temperatures only.*
From Walsh et al. (2011)

Excessive warming at high-latitudes is thought to stem primarily from an efficient climate feedback structure, whereby surface temperature rises initiate the reduction of snow and ice extent which decrease surface albedo, leading to greater absorption of incoming radiation and further warming (Kaplan and New, 2006; Serreze and Francis, 2006). Recent temperature rises in autumn months have coincided with reductions in September sea ice extent and support the concept of an ice-albedo amplification mechanism (Serreze *et al.*, 2009). However, some studies have suggested that other factors may also contribute to the Arctic amplification process, including ocean and atmospheric circulation patterns, cloud cover and water vapour (Graversen *et al.*, 2008; Screen and Simonds 2010). Regardless of the underlying cause, it is considered highly likely that excessive warming in the Arctic will continue in future (Anisimov *et al.*, 2007; Overland *et al.*, 2008) and coupled atmosphere-ice-ocean climate models predict a rise in mean surface air temperature of 5-8 ºC through the rest of the 21st Century (Figure 2.3; Anisimov *et al.*, 2007; Overland *et al.*, 2011; Vavrus *et al.*, 2012). The greatest increase is expected during winter months because melting sea ice is likely to constrain the rise of surface air temps in summer (Lu and Cai, 2009; Overland *et al.*, 2011; but see Cohen *et al.*, 2012).

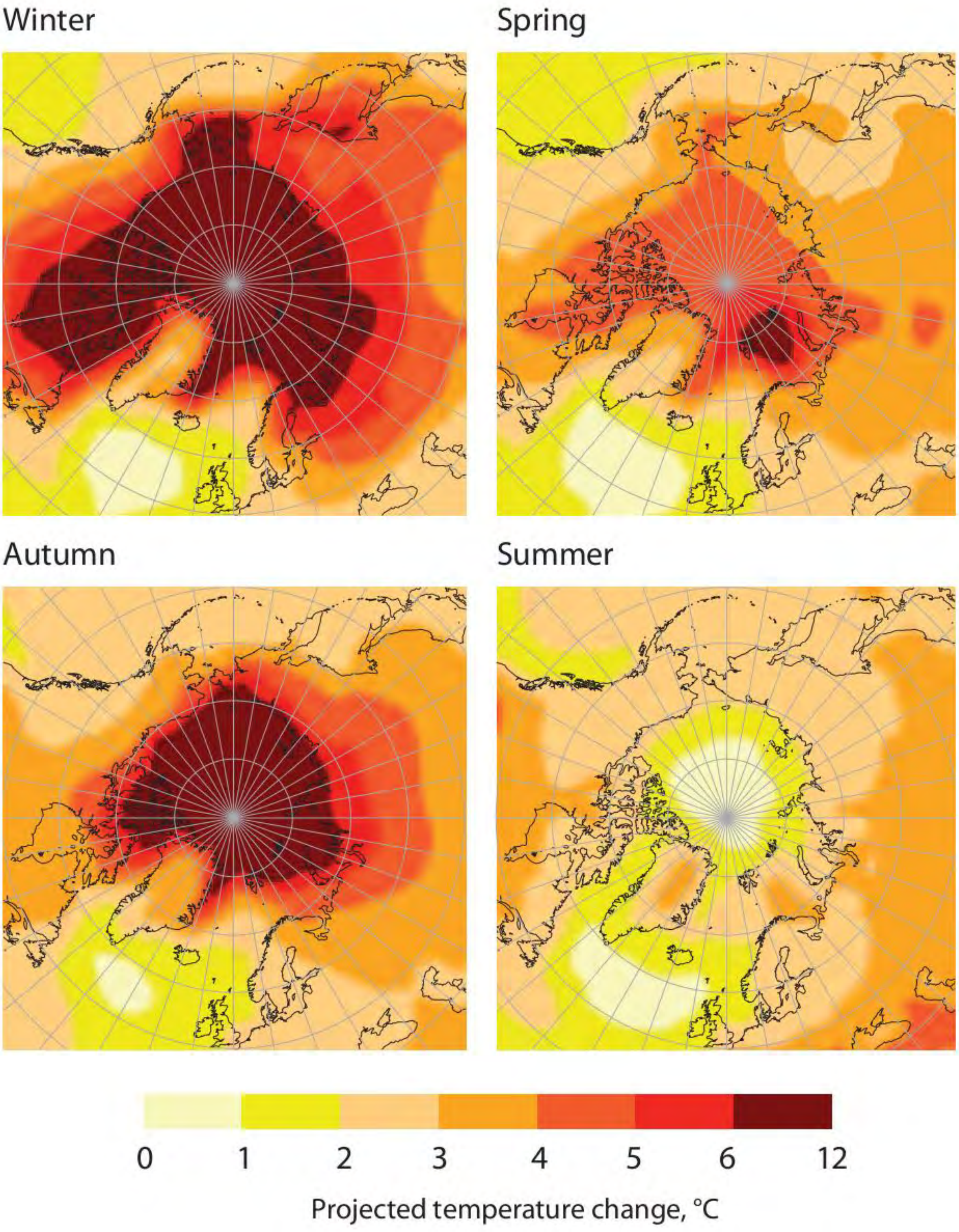


Figure 2.3: *Projected changes of seasonal mean surface air temperature for the period 2070 to 2090. Changes are composited over 14 CMIP3 model projections forced by the IPCC A1B emissions scenario. From Overland et al. (2011)*

Precipitation in the Arctic increased over the past century at about 1% per decade (McBean *et al.*, 2005) although sparse coverage of precipitation gauges in many Arctic regions and large uncertainty in remote-sensing-derived estimates means these findings should be treated with caution (McClelland *et al.*, 2004; Anisimov *et al.*, 2007; Walsh *et al.*, 2011). Such uncertainty complicates predictions of future changes, but in general climate models project increased precipitation over most of the Arctic through the 21st Century (Figure 2.4; Kattsov *et al.*, 2007; Overland *et al.*, 2011; Mahlstein *et al.*, 2012). As with air temperature, the largest relative increases are expected during winter months which could, in combination with warmer conditions, increase the frequency of rain-on-snow events (Rennert *et al.*, 2009).

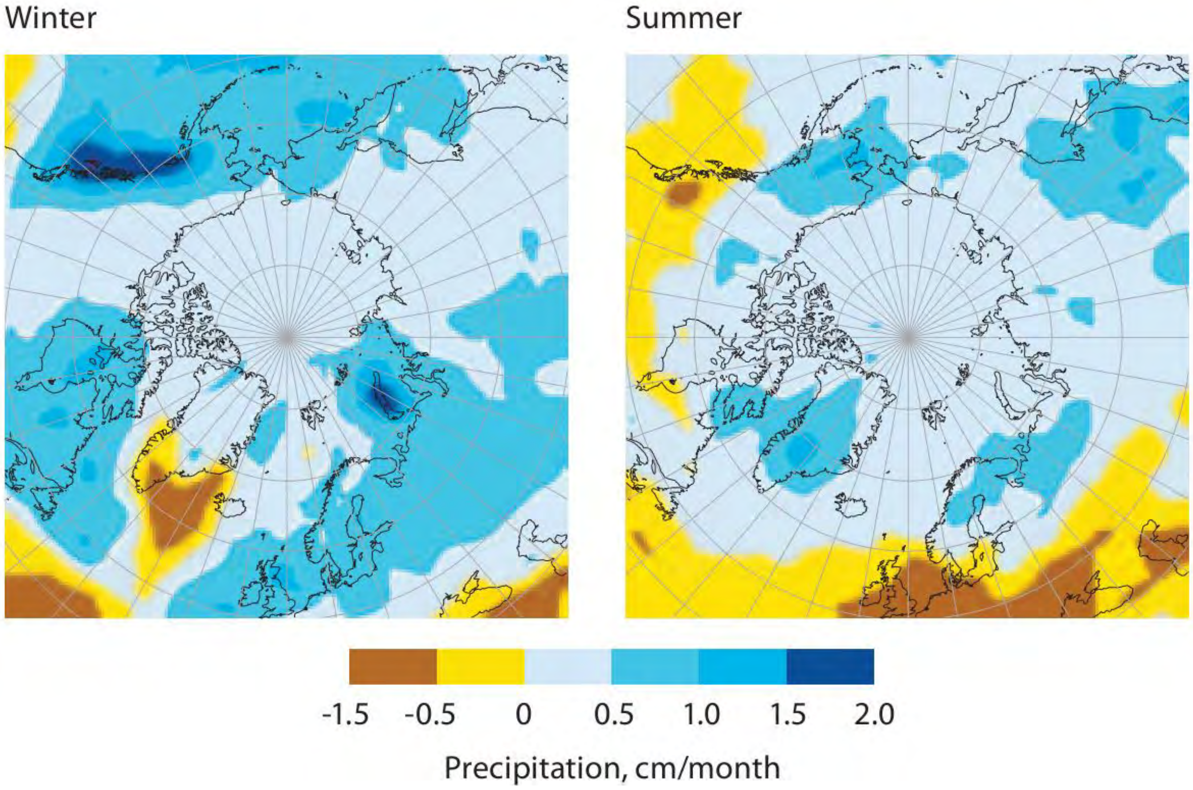


Figure 2.4: *Projected changes for winter (left) and summer (right) precipitation for the period 2070 to 2090. Changes are composited over 14 CMIP3 model projections forced by the IPCC A1B emissions scenario. From Overland et al. (2011)*

2.2.2 Cryospheric processes

The Arctic contains a large number of glaciers which together with the Greenland Ice Sheet cover an area of more than 400,000 km² (Sharp *et al.*, 2011). Glacier thinning in many parts of the world has accelerated during recent years (Zemp *et al.*, 2009), and this is also the case for much of the Arctic (Figure 2.5; Lemke *et al.*, 2007; James *et al.*, 2012). Cryospheric processes are driven primarily by temperature and precipitation (Walsh *et al.*, 2011) and therefore future changes in atmospheric forcing will have implications for regional Arctic mass balance, snow cover and permafrost systems. Although there are few long-term high-latitude glacier observations (Zemp *et al.*, 2009), most monitored glaciers retreated throughout the last century following the end of the Little Ice Age (Sharp *et al.*, 2011). Simulations of glacier response to future climate change show large declines in ice volume, and in some cases complete glacier loss, by the end of the 21st Century (Lemke *et al.*, 2007; Radic and Hock, 2011; Sharp *et al.*, 2011).

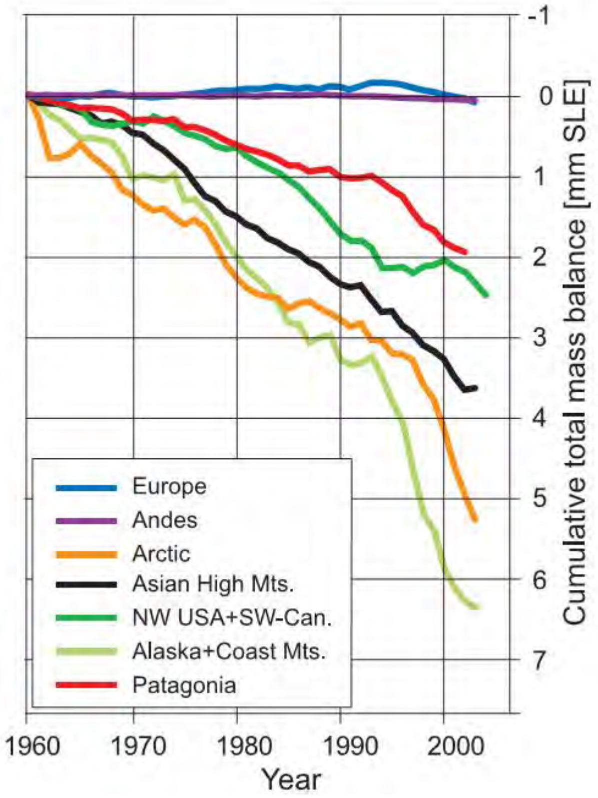


Figure 2.5: *Cumulative total mass balances (sea-level equivalent) of glaciers and ice caps for large regions From Lemke et al. (2007)*

At present, almost 50% of land surface in the Northern Hemisphere is covered by snow during midwinter, of which the majority is located at high-latitudes (Callaghan *et al.*, 2011c). On an annual basis, snow cover has declined by approximately 10% in the last quarter of the 20th Century (Sereze *et al.*, 2000) with the greatest reductions in spring and summer months (Lemke *et al.*, 2007). Despite forecasts for increased precipitation in the Arctic, model projections suggest a future decrease in seasonal snow cover and an increase in the number of snow-free days each year (Anisimov *et al.*, 2007; Callaghan *et al.*, 2011c). Reductions in snowpack coverage, duration and density are likely to amplify the expression of climate warming below the ground surface because snow is an efficient insulator (Stieglitz *et al.*, 2003). Recent decreases in snow cover have accompanied increases in ground temperatures of 0.5 to 2 ºC at almost all permafrost monitoring sites across the Arctic (Callaghan *et al.*, 2011a; Etzelmüller *et al.*, 2011). Moreover, an increase in summer rainfall is likely to be associated with increased soil thawing (Wright *et al.*, 2009). Approximately 24% of the land surface north of the equator is occupied by frozen soils (Zhang *et al.*, 1999). Model projections indicate that the spatial extent of near-surface permafrost could be reduced by up to 90% be the end of the 21st Century (Figure 2.6; Lawrence and Slater, 2005). Melt-induced changes in the structure and distribution of permafrost systems could support the development of deeper flow pathways, increase hydrological connectivity between supra- and sub-permafrost groundwater reservoirs, and enhance rates of geochemical weathering in active soil layers (Walsh *et al.*, 2005; White *et al.*, 2007).

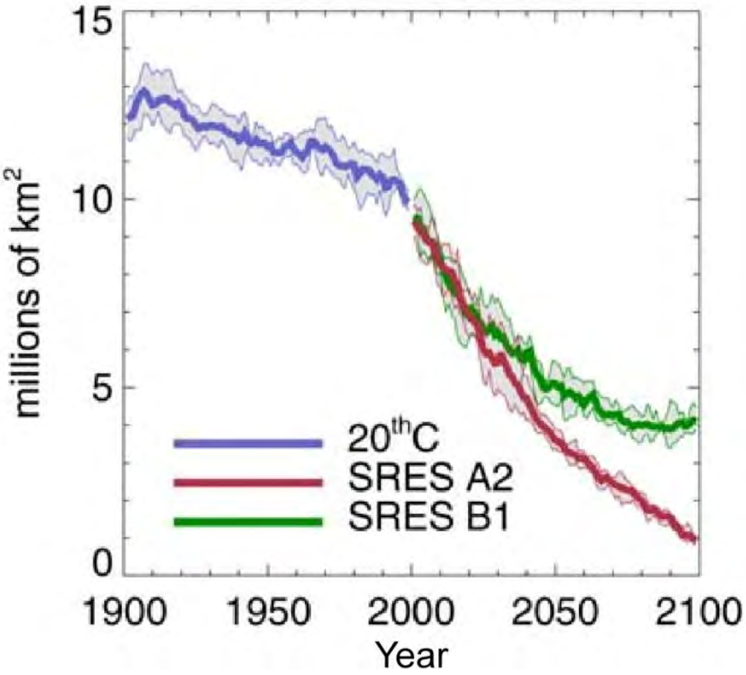


***Figure 2.6:** Observations of global permafrost area, excluding Antarctica and Greenland, during the 20th Century (blue line) and model simulations for the remainder of the 21st Century (red and green lines). Shaded areas show variability around the mean.*

From Lawrence and Slater (2005)

2.2.3 Summary of environmental change in the Arctic

The existing literature demonstrates that the Arctic region is highly sensitive to the effects of global climate change because variations in temperature and precipitation are amplified at high latitudes. In turn, these exert strong controls on regional glacier mass balance, snow cover and permafrost systems. The following sections consider how these changes will affect Arctic river ecosystems. An interdisciplinary approach is taken to establish the chain of processes that link climate and cryospheric systems to local basin water sourcing, river habitat conditions and aquatic biota (*sensu* Hannah *et al.*, 2007). Key areas are identified where more detailed information is required to understand the response of these Arctic river systems to climate change.

2.3 Hydroecological response to environmental change in the Arctic

2.3.1 Hydrological processes and water sourcing in Arctic river basins

Changes in climate are expected to exert strong controls on Arctic river ecosystems because high-latitude rivers originate primarily from frozen water sources (i.e. glaciers, snowpacks, permafrost) that are sensitive to changes in atmospheric processes (Anisimov *et al.*, 2007; Callaghan *et al.*, 2011b). Annual runoff from many highly glacierized high-latitude river basins has risen since 1950 in response to increased glacier ablation (Hinzman *et al.*, 2005) and summer melt seasons have expanded with earlier break-up dates and later freeze-up dates (Lemke *et al.*, 2007). These trends were most pronounced after 1990, coinciding with increases in mean Arctic surface air temperature and precipitation (Shiklomanov *et al.*, 2003; McBean *et al.*, 2005). In future, glacier meltwater production is likely to decrease over inter-annual timescales as glaciers retreat under a warming climate (Barnett *et al.*, 2005), although this may be preceded by a peak in runoff in response to increased energy inputs and enhanced ablation (Milner *et al.*, 2009). Increased winter precipitation and associated snow accumulation may generate higher spring flows (White *et al.*, 2007), although these could be attenuated by deeper active soil layers which promote infiltration of meltwater and thereby buffer peaks in flow regimes (Prowse *et al.*, 2006). Additionally, some evidence now indicates that permafrost wasting has resulted in increased groundwater discharge to Arctic rivers in recent decades (Smith *et al.*, 2007; Walvoord and Streigl, 2007). A greater future likelihood of perennial winter flows in Arctic streams is likely in areas where permafrost melting is substantial (Huryn *et al.*, 2005). Furthermore, future flow regimes may shift from nival- to pluvial-dominated in areas where rainfall events become more frequent and the fraction of precipitation occurring as rain increases (Stewart, 2009).

Predicting future Arctic river flow regimes requires a comprehensive understanding of the hydrological processes governing river flow dynamics. The principal controls on meltwater generation and major runoff pathways of glacier-fed rivers in the proglacial areas of Arctic regions have been studied relatively widely (Hodgkins, 1997; Rutter *et al.*, 2011). However, hydrological research in many Arctic glacierized basins has focussed typically on glacial processes (Hodgkins *et al.*, 2009) and fewer studies have investigated the wider basin context where seasonal snowpacks, buried ice, rainfall-runoff and groundwater systems also contribute to bulk river discharge (Figure 2.7; Milner *et al.*, 2009; Cooper *et al.*, 2011). Additionally, the processes of flow generation can differ between glacial- and nival-dominated rivers (Marsh and Woo, 1981; Kostrzewski *et al.*, 1989; Fleming *et al.*, 2006), indicating that non-glacierized river basins may be important indicators of future flow regime changes in regions where glacier recession is ongoing.

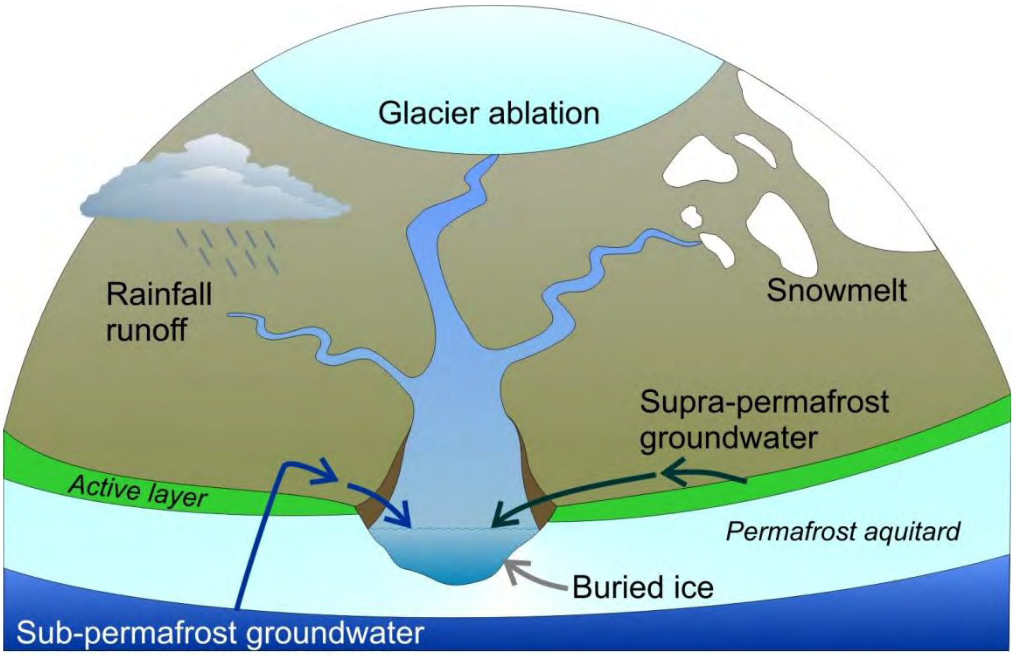


Figure 2.7: *Potential water sources contributing to river flow within an Arctic glacierized basin*

Consequently, there remains a need for more detailed investigations of the key basin water sources that control flow dynamics in Arctic environments (e.g. Woo, 1983). Understanding how these vary temporally over seasonal and interannual timescales will advance our functional understanding of basin hydrological processes in Arctic regions and permit more robust predictions of hydrological responses to broad-scale environmental change in these river systems (Pohl *et al.*, 2007). Variations in the seasonal contribution of water sources to river flow in a glacierised Arctic river basin (Figure 2.8a) are hypothesised to comprise of (i) early melt season flow contributions from snowpack melting, (ii) seasonal progression from snowmelt-dominated to glacier runoff-dominated river flow, and (iii) continuous low discharge from groundwater stores with aquifer recharge in summer. During the rest of the 21st Century, warmer climatic conditions may stimulate glacier ablation and increase river discharge during summer months (Figure 2.8b; Milner *et al.*, 2009). However, following loss of glacier ice, river discharge is likely to decrease in the long term (Figure 2.8c; Barnett *et al.*, 2005).

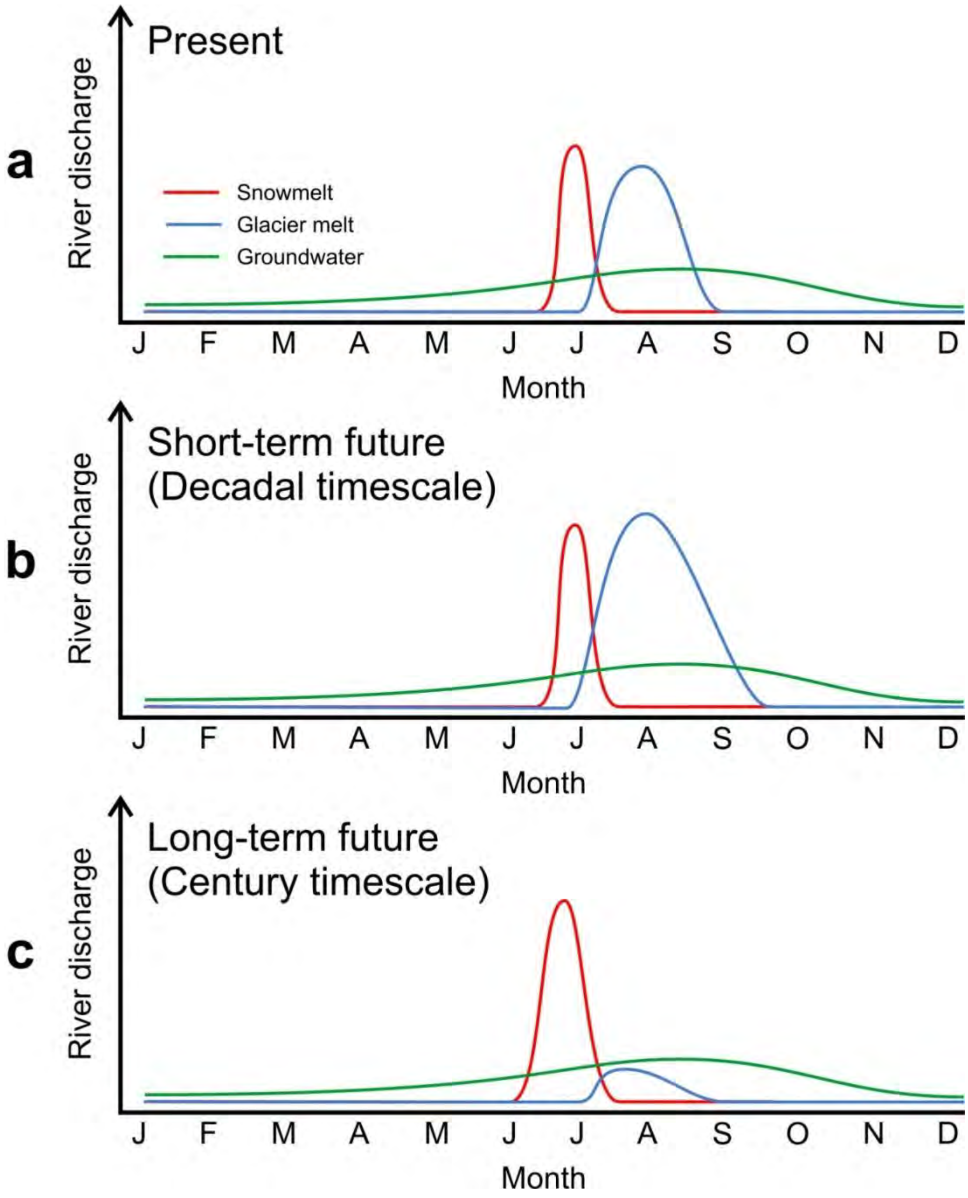


Figure 2.8: *Hypothesized present and future seasonal timing and magnitude of relative levels of water source contribution to river discharge in an Arctic glacierized basin*

Solute loads in many Arctic rivers are derived principally from atmospheric deposition and rock-water interactions in sub-glacial and ice-marginal areas (Wadham *et al.*, 2001; Hodson *et al.*, 2002). Routing of snow and glacial meltwaters over or through recently exposed geochemically-active morainic material leads to river runoff characterised by high concentrations of Ca^{2+} and HCO_3^- (Hodson *et al.*, 2002; Rutter *et al.*, 2011). Concentrations of SO_4^{2-} can also be elevated in meltwater from glaciers with distributed sub-glacial drainage structures due to sulphide oxidation at the glacier bed (Wadham *et al.*, 1998; Krawczyk and Bartoszewski, 2008). In proglacial areas, increased water residence times in older sediments promote silicate weathering and increase the acquisition of solutes from crustal sources in active soil layers (Anderson *et al.*, 2000; Dragon and Marciniak, 2010), thus imparting a hydrochemical signature on shallow groundwater discharge that contrasts with that of glacier and snowpack meltwater. Similarly, the isotopic composition of bulk runoff may also reflect variations in the geographical source of bulk river flow (Theakstone, 2003; Liu *et al.*, 2004) because high rates of evaporation in proglacial areas during summer months lead to the progressive enrichment of heavy-isotope content in active layer soil waters relative to that of precipitation and snowmelt (Cooper *et al.*, 2002; Gibson, 2002).

Consequently, the hydrochemical and isotopic signatures of Arctic river waters may provide insight into the hydrological processes that operate within Arctic river basins. This may allow for the delineation of bulk hydrographs into individual source components (Laudon and Slaymaker, 1997) because spatiotemporal differences in hydrochemical and isotopic characteristics can reflect variations in the routing of water from different basin storage zones (Tranter *et al.*, 1996; Malard *et al.*, 1999). Such information is required to advance our functional understanding of basin hydrological processes in high-latitude regions, and to infer potential changes in basin water storage contributions under a changing climate. Previous studies have demonstrated the effectiveness of these approaches in lower-latitude mountainous environments (Sueker *et al.*, 2000; Brown *et al.*, 2006b; Cable *et al.*, 2011) but their application to Arctic river basins has been relatively under-utilised and limited largely to event-based studies (McNamara *et al.*, 1997 but see Obradovic and Sklash, 1986).

In summary, the literature review of hydrological processes and water sourcing in Arctic river basins indicates that there is currently limited knowledge of the processes that govern river flow dynamics in these systems. Hydrochemical and isotopic characteristics of Arctic river waters may allow for the identification of conceptual basin water sources and offer potential for generating quantitative estimates of seasonal and year-to-year dynamics in water source contributions to bulk flow.

2.3.2 Water temperature dynamics in Arctic river basins

Water temperature shapes the physical, chemical and biological properties of aquatic environments (Caissie, 2006; Webb *et al.*, 2008). A detailed understanding of thermal dynamics in high-latitude rivers is important because one of the most direct impacts of climate change on Arctic river systems could be the modification of water column temperature associated with atmospheric forcing and changes in local hydrology and hydrogeology (White *et al.*, 2007). The role of water temperature as a driver of ecosystem processes is particularly significant in glacierized headwater river basins (Milner *et al.*, 2001; Brown *et al.*, 2005; Cadbury *et al.*, 2008; Roy *et al.*, 2011) due to strong coupling and high sensitivity between atmospheric and in-stream processes (Hannah *et al.*, 2007).

Water source is hypothesised to play an important role in determining river temperature in glacierized catchments. For example, Brown *et al.* (2005) and Brown and Hannah (2008) recorded consistently low water temperature (-0.3 – 2.3 ºC) at a Pyrenean glacier snout and increased temperature (max. 15.4 ºC) at downstream sites, but noted that discontinuities in the longitudinal temperature gradient were introduced by cool thermally-stable groundwater tributary inputs to the river main stem. Similar findings have also been reported by Milner *et al.* (2001) and Cadbury *et al.* (2008) for other sites in alpine Europe and the Southern Alps of New Zealand, respectively. Additionally, while heat budgets of rivers are often dominated by net radiation inputs (Webb and Zhang, 1997; Evans *et al.*, 1998; Hannah *et al.*, 2004; 2008) especially in mountainous environments above the tree-line where little riparian shading occurs, sustained input of precipitation and associated runoff to glacial channels can decouple radiation-water temperature relationships, particularly under overcast skies when short-wave radiation inputs are reduced (Brown & Hannah, 2007; Chikita *et al.*, 2010).

While in-depth examinations of seasonal thermal regimes exist for several high-altitude headwater alpine catchments (Uehlinger *et al.*, 2003; Brown *et al.*, 2005; Cadbury *et al.*, 2008), river thermal dynamics remain relatively understudied in Arctic environments (for exceptions see Irons and Oswood, 1992; Lammers *et al.*, 2007; Chikita *et al.*, 2010). In contrast to alpine environments, high-latitude environments are characterised by stronger seasonality, lower incoming solar radiation, persistently colder air temperatures, and permafrost that underlies the majority of land masses >66 ºN (Power and Power, 1995). Permafrost lenses in the soil prevent deeper groundwater-surface water interactions (Judd and Kling, 2002), thus potentially limiting the capacity of thermally-stable groundwater inputs to buffer variability in temperature regimes (Brown *et al.*, 2005). It follows that differences in the magnitude and timing of incoming solar radiation received at high-latitudes may cause less diurnal variability but greater annual river temperature dynamics compared to those seen in alpine systems.

Future climatic changes are likely to affect two key aspects of high-latitude water temperature dynamics. Firstly, by controlling the timing of break-up and freeze-up dates and thus determining the magnitude and duration of the annual summer flow period (Prowse *et al.*, 2006; Prowse and Brown, 2010), and secondly, by influencing river thermal regimes at seasonal to sub-daily timescales by modifying river discharge and atmospheric energy receipt (Irons and Oswood, 1992; Caissie, 2006; Lammers *et al.*, 2007; Chikita *et al.*, 2010). Strong air-water column temperature relationships observed in lower-latitude regions (Mohseni and Stefan, 1999) imply Arctic rivers may undergo a shift towards warmer thermal regimes under forecasted increases in surface air temperature (Anisimov *et al.*, 2007; Vavrus *et al.*, 2012). In turn, such changes are likely to alter both physicochemical habitat conditions for freshwater biota (Prowse *et al.*, 2006; Hannah *et al.*, 2007; Milner *et al.*, 2009) and also the metabolic rates of ectothermic biota (Brown *et al.*, 2004; Woodward *et al.*, 2009). However, contemporary understanding of the major characteristics and drivers of water temperature dynamics in high-latitude basins is minimal. Further research is required in this subject area to investigate variability in thermal regimes and explore links to controlling hydroclimatological factors in Arctic rivers. Such information would enable a more accurate assessment of the thermal response of these systems to climate change.

2.3.3 Nutrient cycling in Arctic rivers

Future climate warming at high-latitudes and associated changes in hydrological processes are likely to result in large modifications to freshwater and nutrient budgets due to strong land-water linkages (Judd and Kling, 2002) and the vast quantity of organic matter stored in Arctic tundra ecosystems (Tarnocai *et al.*, 2009; Keuper *et al.*, 2012). Stocks of terrestrial carbon in the Arctic account for around one third of the global total (McGuire *et al.*, 1995) and the amount of dissolved organic carbon (DOC) delivered by rivers each year to the Arctic Ocean is estimated at 27.7 Tg DOC year^{-1}, exceeding that of any other world ocean and therefore exerting a strong control on food availability for marine organisms (Lobbes *et al.*, 2000). However, much of this DOC is refractory after the spring freshet, making C a potentially limiting nutrient for biota in riverine or oceanic environments for a large proportion of each year (Dittmar and Kattner, 2003; Amon and Meon, 2004; Holmes *et al.*, 2008). Similarly, inorganic nutrient concentrations in Arctic rivers are currently among the lowest in the world (Dittmar and Kattner, 2003) and consequently autochthonous primary production in these systems is low (Cauwet and Siderov, 1996; Tank and Dodds, 2003).

Under future warming scenarios a rise in soil temperature is likely to stimulate mineralization of stored organic nitrogen (Jones *et al.*, 2005) and a thickening of the seasonal active layer may increase the pool sizes of nitrogen and carbon that could potentially be flushed into rivers (Frey and Smith, 2005; Petrone *et al.*, 2006). Although uncertainty remains, accelerated glacier wasting and permafrost degradation could also increase the proportion of labile DOC in Arctic rivers (Hood *et al.*, 2009; O'Donnell *et al.*, 2010). Increasing ambient nutrient availability is likely to enhance primary productivity and microbial respiration in these systems (Lock *et al.*, 1990; Hershey *et al.*, 1997), and may also release inorganic N assimilation from DOC limitation (Dodds *et al.*, 2004; Johnson *et al.*, 2009). Moreover, a progressive shift towards groundwater-dominated flow regimes is likely to be linked to the modification of physicochemical habitat conditions in high-latitude river catchments (Huryn *et al.*, 2005). Enhanced groundwater contributions are associated with more stable hydrological regimes, warmer water temperature and lower suspended sediment loads than glacier-fed rivers (Ward *et al.*, 1999; Parker and Huryn, 2006; Blaen *et al.*, *in press*), thus providing more suitable habitat conditions for growth and development than meltwater-dominated systems (Prowse *et al.*, 2006).

The dual effects of nutrient loading and improved habitat quality for organisms in high-latitude rivers may result in substantial changes in particular ecosystem processes associated with increased primary production and microbial respiration (Lecerf and Richardson, 2010). In particular, variation in nutrient uptake rates in response to climate change has important management implications for reducing nutrient loading and regulating water quality in downstream areas (Alexander *et al.*, 2007). Small headwater rivers play a key role in this function because they comprise over 80% of total drainage network length and have high processing rates relative to larger channels downstream (Alexander *et al.*, 2000; Peterson *et al.*, 2001; Craig *et al.*, 2008). While there are few studies on nutrient uptake in headwater Arctic rivers (but see Wollheim *et al.*, 2001; Scott *et al.*, 2010), the extreme cold, highly unstable channels, and limited pool of bioavailable nutrients (Friberg *et al.*, 2001; Hodson *et al.*, 2002; Huryn *et al.*, 2005; Petrone *et al.*, 2006; Holmes *et al.*, 2008; Blaen *et al.*, *in press*) that characterise many meltwater-dominated high-latitude systems suggest that contemporary rates of nutrient uptake are low in these environments (Scott *et al.*, 2010). In the context of projected warming in polar regions, increased autotrophy and microbial activity may lead to a subsequent intensification in nutrient uptake and rates of retention in Arctic rivers. Moreover, these processes could be further enhanced by the deepening of thaw bulbs below main river channels associated with permafrost melting (Zarnetske *et al.*, 2007; Brosten *et al.*, 2009). Hyporheic transient storage areas, where waters are stationary relative to the main channel, are thought to be important zones of biogeochemical processing, but may be limited in depth by permafrost layers in high-latitude areas (Edwardson *et al.*, 2003; Greenwald *et al.*, 2008; Merck *et al.*, 2012).

Consequently, there is a current paucity of information regarding nutrient cycling in Arctic rivers. At present, both rates and controlling factors of nutrient uptake in these systems remain unclear. There is a need for research to understand how future changes in climate and permafrost systems may affect river nutrient fluxes and dynamics in high-latitude environments.

2.3.4 Benthic macroinvertebrate response to hydrological change in Arctic rivers

The Arctic contains some of the most inhospitable habitat available for freshwater biota due to severe environmental conditions imposed by high seasonality, intermittent flow regimes, short summers and long, cold winters (Lods-Crozet *et al.*, 2001; Prowse *et al.*, 2006; Avila-Jimenez *et al.*, 2007; Danks, 2007). Ongoing and future hydrological changes in high-latitude rivers are expected to have strong impacts on Arctic freshwater biodiversity (Prowse *et al.*, 2006; Wrona *et al.*, 2006; Vincent *et al.*, 2011) due to the modification of aquatic habitat conditions and thus community dynamics, as observed previously in alpine regions (Brown *et al.*, 2007a). These systems may be useful and sensitive indicators of environmental change because community assembly and succession in such harsh environments is strongly deterministic due to the overriding dominance of physical factors (Milner *et al.*, 2008; Brown and Milner, 2012). However, freshwater biota in many Arctic regions has been relatively understudied in comparison to lower-latitude areas, and the potential response of these river ecosystems to future hydrological change is not well understood.

Glacier-fed rivers in high-latitude regions are physically-dynamic systems that exert strong selective forces directly on aquatic biota through a combination of cold water temperature, unstable channel morphologies, sediment abrasion, ice scour and seasonal freezing (Milner *et al.*, 2001; Wrona *et al.*, 2006; Huryn *et al.*, 2005). In contrast, rivers derived from non-glacial sources (e.g. groundwater) are characterised typically by warmer water temperature, greater electrical conductivity (EC), lower suspended sediment concentration (SSC) and greater channel stability (Ward, 1994; Prowse *et al.*, 2006; Blaen *et al. in press*). Many non-glacier-fed Arctic rivers support increased abundance and diversity within macroinvertebrate communities relative to glacier-fed rivers (Craig and McCart, 1975; Parker and Huryn, 2011), particularly where perennial flows from springs provide overwintering refugia (Ward, 1999). Such research indicates that rivers dominated by groundwater flow are important for Arctic river biodiversity because they often contain more favourable habitat conditions than rivers sourced primarily from snow and glacier meltwater (Ward, 1994; Huryn *et al.*, 2005). The dominance of these primary river water sources can vary in space (i.e. between rivers) and also in time association with glacier retreat, changes in snowpack volume, or variation in groundwater discharge (Brown *et al.*, 2003; Milner *et al.*, 2009).

Previous studies from lower-latitude alpine regions have quantified links between water source contributions (i.e. meltwater, groundwater), habitat conditions and benthic macroinvertebrate communities (Brown *et al.*, 2007b; Brown *et al.*, 2009). Decreases in the proportion of river flow comprised of meltwater are associated generally with increases in aquatic macroinvertebrate abundance and diversity, although some cold-stenotherms may be vulnerable to meltwater reduction (Brown *et al.*, 2007a; Jacobsen *et al.*, 2012). Therefore, despite their low alpha diversity, cold glacier-fed headwater rivers in Arctic areas may harbour unique benthic macroinvertebrate taxa that are important contributors to regional biodiversity (Meyer *et al.*, 2007). Moreover, benthic macroinvertebrate taxa represent an important food source for higher trophic organisms (Wipfli and Baxter, 2010) and therefore future changes in biodiversity could have broad ramifications for wider Arctic food webs (Parker & Huryn, 2006; Prowse *et al.*, 2006). From a conservation viewpoint, understanding the hydrological drivers of river habitat conditions and biodiversity could assist in the development of models to predict the response of these freshwater ecosystems to future warming (Milner *et al.*, 2009; Ormerod, 2009), particularly if water source contribution ‗optima' can be identified for individual species (*cf.* Brown *et al.*, 2007b). However, such information is currently lacking for high-latitude river basins, despite the potential for rapid environmental change in northern regions (McBean *et al.*, 2005).

To conclude, the literature review indicates that at present there is minimal understanding of how hydrological changes in high-latitude river basins will impact aquatic benthic macroinvertebrate communities. There is now a need to quantify linkages between dynamic water source contributions, river habitat conditions and freshwater biota in Arctic river basins to achieve an integrated understanding of how these systems might respond to climate change.

2.4 Summary of key research gaps

The literature review has identified four research gaps that warrant further investigation. These were first presented in section 1.1 and are outlined again below:

1) There is currently inadequate knowledge of the hydrological processes and key water sources that govern river flow dynamics in high-latitude river systems (Hodgkins *et al.*, 2009; Cooper *et al.*, 2011). This research gap is important because such knowledge is necessary to underpin accurate predictions of future flow regimes in Arctic river basins.

2) Very few studies have examined water temperature regimes in high-latitude basins and little is known of the major drivers of thermal variability in these systems (Irons and Oswood, 1992; Lammers *et al.*, 2007; Chikita *et al.*, 2010). Such knowledge is required to develop a more comprehensive understanding of how climate forcing will alter water temperature dynamics in Arctic rivers.

3) The processes that control rates of nutrient uptake in Arctic rivers are not well understood, despite concerns that permafrost degradation in high-latitude regions could increase nutrient loading to freshwater environments (Petrone *et al.*, 2006; Holmes *et al.*, 2008). There is a need for research to understand how future changes in climate may affect aquatic nutrient fluxes and dynamics in these systems.

4) There is minimal understanding of the links between water source contributions, aquatic habitat conditions and benthic macroinvertebrates in Arctic river basins (Prowse *et al.*, 2006; Wrona *et al.*, 2006; Vincent *et al.*, 2011). This research gap is important because freshwater taxa in these systems may be useful and sensitive indicators of high-latitude environmental change.

2.5 Research objectives

This thesis addresses four objectives derived from the research gaps identified by the literature review. The interrelationships between these conceptual domains are shown (Figure 2.9).

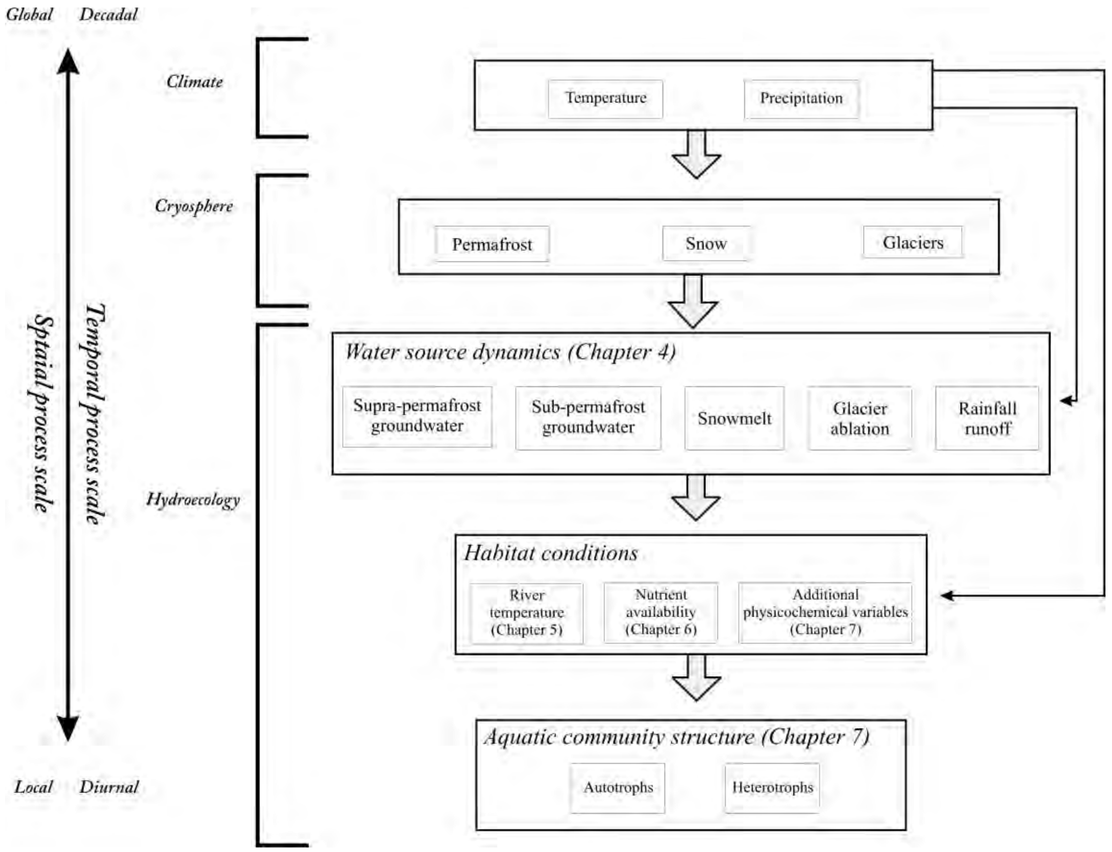


Figure 2.9: *Conceptual model of climate-cryosphere-hydroecology linkages in Arctic river basins operating at different scales. Arrows demonstrate theoretical relationships between components.*

Objective 1: Hydrological processes and water sources in Arctic river basins

The first thesis objective was to use hydrochemical and isotopic data to examine identify conceptual water sources, and quantify seasonal and year-to-year variability in water source contributions to river flow in Arctic river basins. This research objective is addressed in Chapter 4.

Objective 2: Water temperature dynamics in Arctic rivers

The second thesis objective was to use a high-resolution temperature monitoring network to characterise spatiotemporal water temperature dynamics in Arctic rivers and assess the primary drivers of thermal variability in these systems. This research objective is addressed in Chapter 5.

Objective 3: Nutrient cycling in Arctic river systems

The third thesis objective was to adopt an experimental approach to identify and quantify the processes that control rates of nutrient uptake in Arctic rivers to assess how these might vary under a changing climate. This research objective is addressed in Chapter 6.

Objective 4: Ecological responses to hydrological change

The fourth thesis objective was to quantify relationships between water source contributions, physicochemical habitat conditions and biotic community structure to achieve a more comprehensive understanding of hydroecological process-linkages in Arctic rivers. This research objective is addressed in Chapter 7.

2.6 Chapter summary

This chapter has provided an overview of the effects of climate change on Arctic river systems and reviewed previous research on water sourcing and hydrological processes, river temperature dynamics, nutrient cycling and uptake, and drivers of aquatic community structure in high-latitude basins. Four key gaps for investigation have been identified and corresponding research objectives identified to address these. The research design and specific details of the study area, data and methods used in the thesis are presented in the next chapter.

CHAPTER 3

RESEARCH DESIGN, STUDY AREA, DATA AND METHODS

3.1 Introduction

The design of the thesis is documented in this chapter which describes how the research objectives outlined in section 2.5 link together. The selection of the study area is justified and information is provided on landscape characteristics (climate, geology and vegetation), river basin selection, general sampling methodologies, additional data sources and data processing techniques. More detailed information on specific techniques and data analyses are contained in later research chapters.

3.2 Research design

The research conducted in this thesis is contained primarily within Chapters 4-7. Hydrochemical and isotopic data is utilised to identify the processes that govern flow regimes in Arctic river basins and assess seasonal and year-to-year variability in dynamic basin water source proportions (Chapter 4), hence addressing the first research gap (section 2.4). Following this, a high-resolution network of water temperature sensors is used to examine thermal variability and infer the major drivers of water temperature in Arctic rivers (Chapter 5), thus addressing the second research gap. Research in Chapter 6 draws on that of Chapter 4 by taking an experimental approach to investigate the impacts of shifting water sources on rates of nutrient uptake in Arctic rivers to address the third research gap. Similarly, hydrological and ecological datasets are analysed in combination in Chapter 7 to assess the influence of water source dynamics on aquatic habitat conditions and benthic macroinvertebrate communities in Arctic rivers, thus addressing the final research gap. The thesis is concluded by synthesising the findings of the research and discussing potential areas for further research (Chapter 8).

3.3 Study area

To address the research objectives of the thesis (section 2.5), it was necessary to work in a region of the Arctic containing several rivers supplied by different basin water sources. The Brøgger peninsula in the Kongsfjorden area of north-west Spitsbergen, the largest island of the Svalbard archipelago (Figure 3.1), was selected for its suitability because it:

a) contains a number of small basins of varying glacierization within close proximity to one another with rivers supplied by several basin water sources (i.e. snow and glacier meltwater, groundwater, rainfall-runoff), thus facilitating an intensive study design.
b) is readily accessible by international air travel and more locally by foot and boat.
c) accommodates the Ny Ålesund international research village from where a large volume of prior research has been conducted on the local environment, including studies of geology (Hjelle, 1993; Harland et al, 1997), hydrology (Hodson *et al.*, 1998a; Hodson *et al.*, 1998b, Killingtveit *et al.*, 2003), glaciology (Hambrey *et al.*, 2005; Barrand *et al.*, 2010), biogeochemistry (Hodson *et al.*, 2002; Telling *et al.*, 2011) and aquatic ecology (Lods-Crozet *et al.*, 2001; Marziali *et al.*, 2009). These previous studies provide context and information to support the research conducted in this thesis.

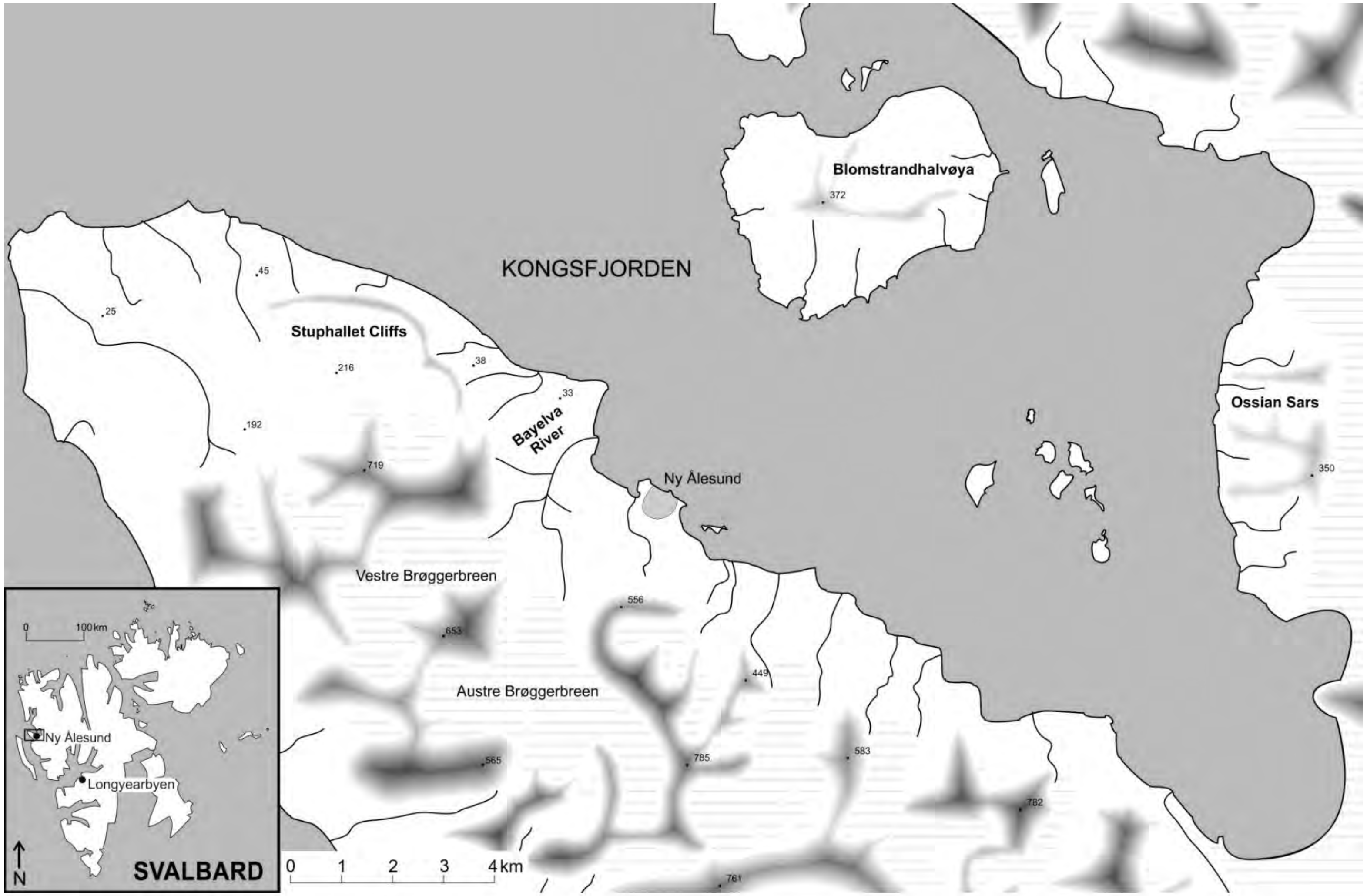


Figure 3.1: *Map of the Brøgger peninsula and wider Kongsfjorden area showing approximate river courses, relief (shaded), spot height measurements in meters asl, and glacier cover (dashed areas).*

3.3.1 Climate of the Brøgger peninsula

Meteorological data have been collected continuously in Ny Ålesund since 1974 by the Norwegian Meteorological Institute. The mean annual air temperature is -6.3 ºC with the warmest mean temperature occurring in July (4.9 ºC; eKlima, 2012). Precipitation is approximately 400 mm yr^{-1} and falls mainly as snow that covers the ground for at least 230 days yr^{-1} (Cannone *et al.*, 2004). More detailed information on atmospheric energy fluxes and recent climatic variability in Svalbard can be found in Førland and Hanssen-Bauer (2003), Westermann *et al.* (2009) and Førland *et al.* (2011).

3.3.2 Geology of the Brøgger peninsula

The majority of the Brøgger peninsula is comprised of calcitic and dolomitic formations (Dallmann *et al.*, 2002). The Ny Ålesund area is typified by sedimentary rocks of the Middle and Upper Carboniferous and Permian ages which include limestone and red sandstones (Hjelle, 1993). Tertiary and Quaternary deposits, including shales and coal seams, are also present to the south and west of Ny Ålesund. Further information on the geology of the Brøgger peninsula and wider Svalbard area can be found in Hjelle (1993) and Harland *et al.* (1997).

3.3.3 Vegetation of the Brøgger peninsula

Vegetation in Svalbard is sparse and trees are absent throughout the entire archipelago, although vascular plant species richness is greater than in other areas at similar latitudes owing to the warming influence of the Gulf Stream system on the local climate (Cooper *et al.*, 2004). Plant communities on the Brøgger peninsula are dominated by slow-growing polar willow (*Salix polaris*), mountain avens (*Dryas octopetala*) and purple saxifrage (*Saxifraga oppositifolia*) which colonise thin soil layers in proglacial and coastal areas (Muraoka *et al.*, 2008; Nakatsubo *et al.*, 2008). Moss layers up to 10 cm in thickness are common in poorly drained areas and species include marsh bryum (*Bryum pseudotriquetum*), ribbed bog moss (*Aulacomnium palustre*) and tomentypnum moss (*Tomentypnum nitens*) (Høj *et al.*, 2006). Therefore, riparian canopy cover and associated shading are absent at all rivers on the peninsula, and allochthonous inputs of particulate organic matter are restricted to those derived from aeolian deposition.

3.4 River basin selection

River basins were selected to address the research objectives (section 2.5) on the basis of their size, proximity to Ny Ålesund, and whether they were considered to be representative of similar basins in the area. Not all selected basins were utilised in each research chapter given differences in the methodologies used throughout the thesis. However, many river basins were employed to fulfil more than one thesis objective. River basins employed in Chapters 4, 5 and 7 were located in two major study areas, the Bayelva and Stuphallet cliffs areas, which are described in more detail in sections 3.4.1 and 3.4.2. River basins used in Chapter 6 were distributed more widely around the Kongsfjorden area. In total, 23 sites were selected (Table 3.1). The common data period was July 2010 to July 2012.

Table 3.1: *Overview of river sites. For site codes, "V" and "A" denote sites draining Vestre Brøggerbreen and Austre Brøggerbreen, respectively, "N" denotes sites fed by non-glacial sources in the Bayelva area and "S" denotes the Stuphallet cliffs area. Codes X1-X6 and Y1-Y6 denote sites used in Chapter 6 during 2011 and 2012, respectively. Sites utilised in individual chapters are indicated by "x".*

Site code	*Location*	*Primary water source*	*Chapter 4*	*Chapter 5*	*Chapter 6*	*Chapter 7*	*Chapter 8*
V1	Bayelva	Glacier runoff from Vestre Brøggerbreen	x	x		x	x
V2	Bayelva	Glacier runoff from Vestre Brøggerbreen	x	x		x	x
A1	Bayelva	Glacier runoff from Austre Brøggerbreen	x				x
A2	Bayelva	Glacier runoff from Austre Brøggerbreen	x				x
BR	Bayelva	Glacier runoff from Austre and Vestre Brøggerbreen	x	x		x	x
N1	Bayelva	Snowmelt / groundwater		x		x	x
N2	Bayelva	Snowmelt / groundwater		x		x	x
N3	Bayelva	Snowmelt / groundwater		x		x	x
S1	Stuphallet cliffs	Snowmelt / groundwater	x				x
S2	Stuphallet cliffs	Snowmelt / groundwater	x				x
S3	Stuphallet cliffs	Snowmelt / groundwater	x	x		x	x
X1	Bayelva	Snowmelt / groundwater			x		x
X2	Bayelva	Snowmelt / groundwater			x		x
X3	Bayelva	Snowmelt / groundwater			x		x
X4	Bayelva	Snowmelt / groundwater			x		x
X5	Bayelva	Snowmelt / groundwater			x		x
X6	East of Ny Ålesund	Snowmelt / groundwater			x		x
Y1	Blomstrandhalvøya	Snowmelt / groundwater			x		x
Y2	Blomstrandhalvøya	Snowmelt / groundwater			x		x
Y3	Stuphallet cliffs	Snowmelt / groundwater			x		x
Y4	East of Ny Ålesund	Snowmelt / groundwater			x		x
Y5	Bayelva	Snowmelt / groundwater			x		x
Y6	Bayelva	Snowmelt / groundwater			x		x

3.4.1 Bayelva area

The 33.5 km² Bayelva basin contains two glaciers, Austre Brøggerbreen and Vestre Brøggerbreen, which cover 54% of the basin area and are surrounded by steep mountain ridges up to 700 m elevation (Figure 3.1; Figure 3.2; Hagen and Lefauconnier, 1995; ASTER GDEM, 2010). The glaciers are dominated by non-temperate thermal structures and are believed to be predominantly cold-based (Hodgkins, 1997; Hodson *et al.*, 2002; Barrand *et al.*, 2010). Both glaciers have retreated significantly in recent years and the mass-balance of Austre Brøggerbreen has been negative almost consistently since monitoring began in 1967 (Haeberli *et al.*, 2007; Barrand *et al.*, 2010). Seasonally-active braided streams flow across moraine deposits in the upper basin before joining in the lower reaches to form the Bayelva river which discharges into Kongsfjorden after approximately 3 km. Several smaller rivers fed by non-glacial water sources flow across the proglacial area from mountains surrounding the valley. Despite Svalbard being situated within a zone of continuous permafrost up to 500 m deep (Hagen and Lefauconnier, 1995), a shallow groundwater system develops in the Bayelva basin each summer with active soil layers approximately 1 m in depth (Roth and Boike, 2001).

3.4.2 Stuphallet cliffs area

The Stuphallet cliffs are situated approximately 8 km to the west of Ny Ålesund (Figure 3.1). Several small first-order rivers sourced from snowmelt and groundwater originate below the cliffs and flow across an area of poorly drained soils for 1-2 km before discharging into Kongsfjorden through breaks in raised beach ridges (Hjelle, 1993). The cliffs are utilised by seabirds for nesting and breeding during summer months. Abundant vegetation growth occurs on poorly drained soils below the cliffs.

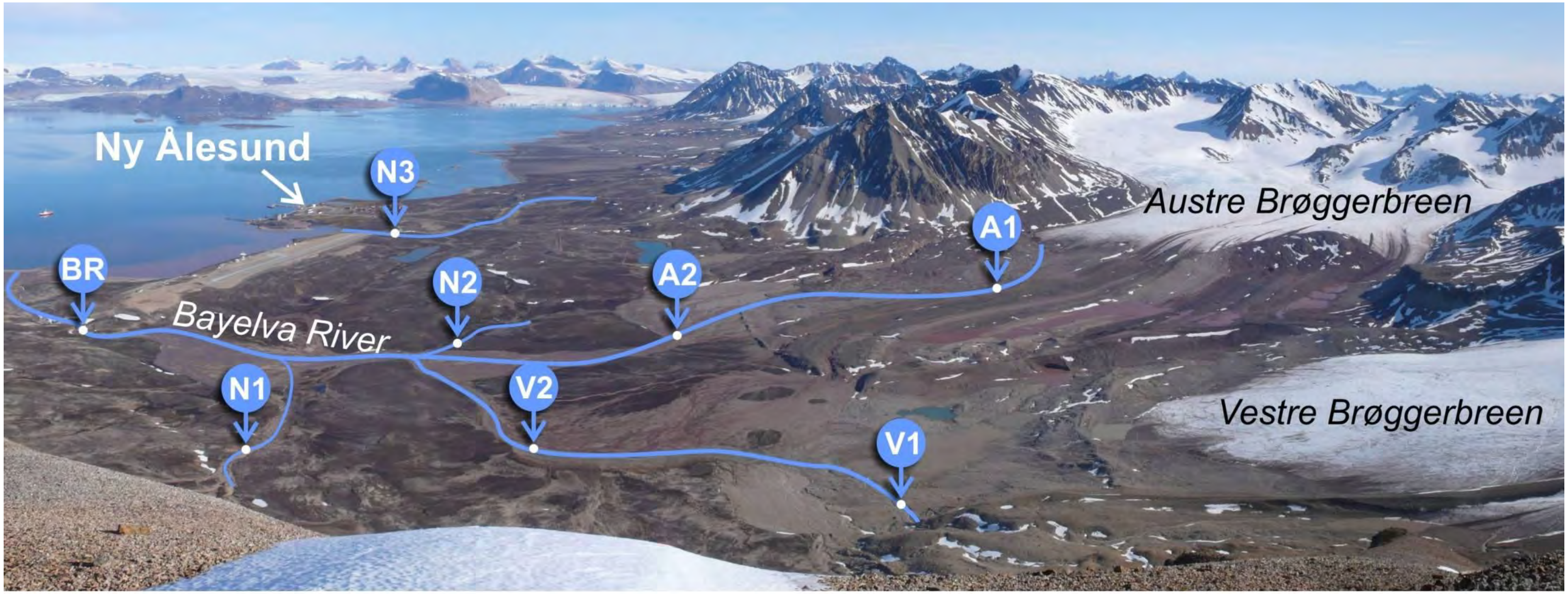


Figure 3.2: *Illustrated photograph of the Bayelva basin area showing approximate river courses, selected sampling sites, Austre and Vestre Brøggerbreen glaciers and the Ny Ålesund research village. See Table 3.1 for further information on sampling sites.*

3.5 General methods

This section describes the sampling and analytical methodologies employed throughout the study. Information on field methods and equipment used to monitor topographical, climatological, hydrological, hydrochemical, isotopic, physicochemical habitat and ecological data are examined in turn and relate to data used in Chapters 4-7. Post-fieldwork data processing and laboratory techniques are discussed in Section 3.7.

3.5.1 Field methods and equipment

3.5.1.1 Topographical and aerial imagery data

A 30 m horizontal resolution and 20 m vertical resolution DEM was derived from the ASTER Global Digital Elevation Model (2010) dataset supplied by the NASA Land Processes Distributed Active Archive Centre. Orthorectified aerial imagery (1 m resolution) was acquired from the NERC Airborne Research and Survey Facility.

3.5.1.2 Climate data

Meteorological data (short-, long- and all-wave incoming radiation (SW↓, LW↓ and AW↓, respectively); air temperature; relative humidity) were measured at 15 min resolution in Ny Ålesund (approximately 1 km and 8 km east of the Bayelva and Stuphallet cliffs, respectively) and supplied by the Alfred Wegener Institute for Polar and Marine Research. SW↓ and LW↓ were measured using a Kipp & Zonen CMP11 pyranometer (typical error ±0.2%) and an EPLAB precision infrared radiometer (typical error ±1%), respectively. Air temperature was measured using a Thies Clima PT100 resistance thermometer (typical error ±0.3 K). Relative humidity was measured with a Vaisala

HMP230 humidity sensor (typical error ±2%). Daily precipitation totals for Ny Ålesund were supplied by the Norwegian Polar Institute.

3.5.1.3 *Hydrological data*

3.5.1.3.1 *River discharge*

River stage was monitored using either Druck PDCR 1840 pressure transducers every 10 s with 15 min averages stored to Campbell Scientific CR10X dataloggers, or Level TROLL 100 (In-Situ Inc.) loggers or TruTrack WT-HT loggers with readings recorded every 15 min. River discharge was calculated over a range of flows at each site using salt dilution gauging (Hudson and Fraser, 2005). Site-specific rating curves were used to provide continuous estimates of river discharge (Appendix A2).

Additional river discharge data were supplied by the Norwegian Water Resources and Energy Directorate. Discharge for the Bayelva River was measured hourly by a compound crump weir and interpolated linearly to 15 min resolution.

3.5.1.3.2 *Water, river bed and local air temperature*

Water and river bed temperature (0.05, 0.20 and 0.40 m depth) was measured using a combination of Tinytag Underwater dataloggers (Gemini Data Loggers (UK) Ltd.) and Campbell Scientific 107 thermistors and CS547A electrical conductivity probes connected to CR10X dataloggers. Local air temperature was monitored using Tinytag Underwater dataloggers housed in radiation shields. Instrumental accuracy was ±0.5 ºC. All internal datalogger clocks were synchronised and checked weekly. Tinytag dataloggers recorded water temperature every 15 minutes. Campbell Scientific instruments scanned every 10 s and 15 min averages were logged. All temperature sensors were crosscalibrated before and after the monitoring period across a temperature range greater than that experienced during fieldwork. Correction factors for individual sensors were computed based on mean values of all sensors during calibration (Hannah *et al.*, 2009). Instrumental drift over each monitoring period was negligible (<0.01 ºC) for all sensors and an order of magnitude less than instrument accuracy.

3.5.1.3.3 Electrical conductivity

Electrical conductivity (EC) was measured using CS547A electrical conductivity probes with either 10 s scans and 15 min averages (Chapters 4 and 5) or discrete measurements every 20 s (Chapter 6) stored to Campbell Scientific CR10X dataloggers. Sensors were inspected regularly and cleaned when necessary to prevent the accumulation of fine sediment within the measuring compartment. Additional spot measurements of EC were obtained using WPA CMD80 and Hanna Instruments 98129 handheld probes.

3.5.1.4 Hydrochemical and isotopic data

This section describes the hydrochemical and isotopic sampling procedures used to understand the major controls governing flows in these Arctic river systems (Chapter 4) and quantify water source contributions to river flow (Chapters 4, 6 and 7).

3.5.1.4.1 Hydrochemical sampling

River water was collected by hand at each river sampling site. Water samples for hydrochemical analysis were filtered immediately through either Whatman 0.45 µm cellulose nitrate (Chapters 4) or 0.7 µm GF/F (Chapter 6) membranes into sterile HDPE bottles ranging from 20 to 125 ml.

Groundwater samples were collected from hillslope seeps in the Bayelva and Stuphallet cliffs areas throughout each study period and processed in a similar manner. Here, groundwater is defined as water residing in the shallow active soil layer that freezes and thaws on a seasonal basis (Roth and Boike, 2001). Discharge from sub-permafrost groundwater was not considered to exert a large influence on surface flows because very few active springs have been documented in the Kongsfjorden area (Haldorsen *et* al., 2010). Snow samples were collected in sterile bags following removal of the upper 5 cm of the snowpack (Brown *et al.*, 2006b). There was a positive bias in snow sample numbers towards the beginning of each field period as the majority of the seasonal snowpack melted by late July. Glacier ice samples were acquired from the lower sections of Austre Brøggerbreen and Vestre Brøggerbreen glaciers in a similar fashion using a clean ice axe to remove the top 10 cm of ice. Safety considerations prevented ice sampling nearer the glacier accumulation zones. Snow and ice samples were allowed to melt in their bags and then processed using the same method as for river water samples. Water samples for hydrochemical analysis were frozen within 8 h of collection and transported to the UK for analysis.

3.5.1.4.2 Isotopic sampling

Unfiltered samples of river water, snow, ice and groundwater were collected in 2 ml glass vials for stable isotope analysis in conjunction with hydrochemical sampling. In addition, rainfall samples were collected using 30 ml glass bottles with a layer of paraffin oil to prevent evaporation prior to storage in glass vials. More frequent samples were acquired during heavy rainfall events. However, sufficient volumes of rain water required for hydrochemical analysis were not achieved due to relatively low precipitation in Svalbard. Water samples for isotopic analysis were stored in a dark container at room temperature and transported to the UK for analysis.

3.5.1.5 Habitat characterisation

3.5.1.5.1 Suspended sediment concentration

Suspended sediment concentration (SSC) was calculated from manual gulp samples (500 ml) filtered onto pre-weighed Whatman 0.45 μm cellulose nitrate membranes, dried at 60 ºC for 48 hrs, and reweighed. Results are expressed in mg l^{-1}. Turbulent river flow at all sites suggested suspended sediments were well-mixed in the water column and that monitoring SSC at the river bank would be representative of the entire river cross-section (Gurnell *et al.*, 1992).

3.5.1.5.2 pH

Spot measurements of pH were obtained using a Hanna Instruments 98129 handheld probe. The instrument was calibrated regularly in pH 4.01 and pH 7.01 buffer solution.

3.5.1.5.3 Channel stability

Channel stability was assessed using the bottom component of the Pfankuch Index (Pfankuch, 1975), where lower values represent greater channel stability. This measure of channel stability was used because it is applicable across a wide discharge range and the assessment methodology is relatively quick to conduct. The index is generated from a visual assessment based on rock angularity and brightness, substrata size and consolidation, scouring and deposition, and the presence of clinging aquatic vegetation. Components are assigned individual scores which are then summed to produce an integrative measure of channel stability.

3.5.1.5.4 Substrate size

Bed sediment material was characterised by measuring the *b*-axis of 100 randomly selected substrate stones and calculating median values (D_{50}) (Brown *et al.*, 2005)

3.5.1.6 Chlorophyll a *concentration*

Samples of periphyton were scrubbed from cobbles using a clean toothbrush, stored in dark bottles, and filtered onto Whatman 0.7 µm GF/F papers within 6 h (Steinman *et al.*, 2006). Each paper was wrapped individually in aluminium foil and frozen until analysis. Cobbles were photographed in the field beside a ruler for scale. Details of chlorophyll *a* extraction and analysis are provided in section 3.7.6.

3.5.1.7 Sediment respiration

Rates of sediment respiration by microbial communities were determined as the change in dissolved oxygen over time (Logue *et al.*, 2004). Following removal of the upper 5 cm of bed sediment, five 1 L HDPE bottles were half-filled with sediment and then topped up with river water. Sediments were sieved to exclude particles >8 mm. Bottles were inverted three times to remove trapped air bubbles and dissolved oxygen concentration was measured using a YSI 95 meter calibrated in the field. Each bottle was sealed and buried in the bed for approximately 24 hours before dissolved oxygen concentration was remeasured. Bottles were transported to the laboratory within 6 h where sediments were dried at 60 ºC for 48 h and weighed. Sediment respiration rates were normalised by sediment weight and reported as µg O_2 h^{-1} kg^{-1} sediment (Logue *et al.*, 2004).

3.5.1.8 Benthic macroinvertebrate sampling

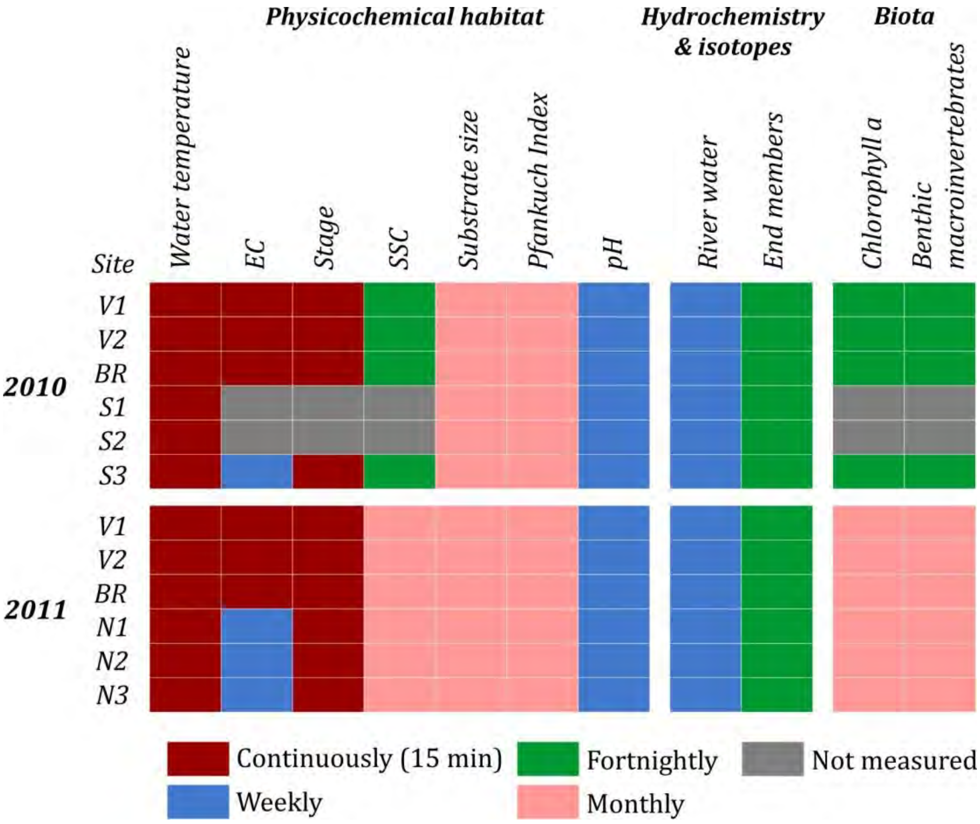
Benthic macroinvertebrate samples were collected from river sites in 2010 and 2011. Samples were collected on the same day from all sites where possible, although logistical constraints meant that samples obtained from sites the Stuphallet cliffs area in 2010 were obtained ±24 h of sites in the Bayelva area.

Five replicate samples of river macroinvertebrates were collected from riffle areas immediately below river gauging equipment using a 0.09 m^2 Surber sampler with a 250 µm mesh on each sampling date. Riffle areas were used because these were accessible at all sampling sites, in contrast with pool areas which were inaccessible in some larger channels. Sampling was conducted moving upriver to prevent habitat disturbance prior to sample acquisition. Individuals were preserved in 70% ethanol and later sorted in the laboratory.

3.6 Sampling resolution

The temporal resolution at which physicochemical habitat, hydrochemical, isotopic and biological variables were sampled at different sites is shown in Table 3.2. Sites used in Chapter 6 are not shown because the methodology of this chapter did not require a multi-day sampling structure.

Table 3.2: *Sampling resolution for research undertaken during 2010 and 2011*



		Physicochemical habitat							Hydrochemistry & isotopes		Biota	
	Site	Water temperature	EC	Stage	SSC	Substrate size	Pfankuch Index	pH	River water	End members	Chlorophyll a	Benthic macroinvertebrates
2010	V1	Continuously (15 min)	Continuously (15 min)	Continuously (15 min)	Fortnightly	Monthly	Monthly	Weekly	Weekly	Fortnightly	Fortnightly	Fortnightly
	V2	Continuously (15 min)	Continuously (15 min)	Continuously (15 min)	Fortnightly	Monthly	Monthly	Weekly	Weekly	Fortnightly	Fortnightly	Fortnightly
	BR	Continuously (15 min)	Continuously (15 min)	Continuously (15 min)	Fortnightly	Monthly	Monthly	Weekly	Weekly	Fortnightly	Fortnightly	Fortnightly
	S1	Continuously (15 min)	Not measured	Not measured	Not measured	Monthly	Monthly	Weekly	Weekly	Fortnightly	Not measured	Not measured
	S2	Continuously (15 min)	Not measured	Not measured	Not measured	Monthly	Monthly	Weekly	Weekly	Fortnightly	Not measured	Not measured
	S3	Continuously (15 min)	Weekly	Continuously (15 min)	Fortnightly	Monthly	Monthly	Weekly	Weekly	Fortnightly	Fortnightly	Fortnightly
2011	V1	Continuously (15 min)	Continuously (15 min)	Continuously (15 min)	Monthly	Monthly	Monthly	Weekly	Weekly	Fortnightly	Monthly	Monthly
	V2	Continuously (15 min)	Continuously (15 min)	Continuously (15 min)	Monthly	Monthly	Monthly	Weekly	Weekly	Fortnightly	Monthly	Monthly
	BR	Continuously (15 min)	Continuously (15 min)	Continuously (15 min)	Monthly	Monthly	Monthly	Weekly	Weekly	Fortnightly	Monthly	Monthly
	N1	Continuously (15 min)	Weekly	Continuously (15 min)	Monthly	Monthly	Monthly	Weekly	Weekly	Fortnightly	Monthly	Monthly
	N2	Continuously (15 min)	Weekly	Continuously (15 min)	Monthly	Monthly	Monthly	Weekly	Weekly	Fortnightly	Monthly	Monthly
	N3	Continuously (15 min)	Weekly	Continuously (15 min)	Monthly	Monthly	Monthly	Weekly	Weekly	Fortnightly	Monthly	Monthly

3.7 Data processing

This section describes post-fieldwork data processing and laboratory techniques used in the thesis.

3.7.1 Catchment delineation and verification

River basins were delineated in a GIS using a raster DEM of the Brøgger peninsula. Basin areas were verified by visual inspection using high-resolution aerial imagery and field observations.

3.7.2 Breaks in time-series data

Periodic instrument errors resulted in occasional breaks in 15-min time-series data. Small breaks (<2 h) in 15 min time-series data were infilled using linear interpolation between data points on each side of the break. Longer breaks (max. 2 days) were corrected using linear regression models between neighbouring sites (Irvine-Fynn *et al.*, 2011).

3.7.3 Stage-discharge relationships

River stage time-series data were converted to discharge estimates using site-specific rating curves derived from salt dilution gauging (Table 3.3; Hudson and Fraser, 2005). Errors in discharge were 10-26% and similar to those reported in other studies of high-latitude hydrology (Hodgkins, 2001; Hodson *et al.*, 2002). Rating curves are displayed in Appendix A2.

Table 3.3: *Stage-Q relationships for river monitoring sites during the 2010 and 2011 melt seasons*

Year	*Site*	*n*	*Stage-Q relationship*	*R²*
2010	V1	25	$Q = 0.007Stage^2 - 0.064Stage - 0.118$	0.90
	V2	15	$Q = 8E^{-08}Stage^{4.943}$	0.76
	S3	9	$Q = 7.560Stage^{2.559}$	0.88
2011	V1	10	$Q = 0.011Stage^{1.701}$	0.74
	V2	8	$Q = 0.586E^{0.052Stage}$	0.79
	N1	9	$Q = 0.026E^{13.80Stage}$	0.74
	N2	9	$Q = 0.00000021Stage^{4.901}$	0.86
	N3	12	$Q = 0.023Stage - 0.229$	0.88

3.7.4 Hydrochemical analysis

Hydrochemical analysis of water samples was undertaken at the University of Birmingham and University of York. Major ion and acetate concentrations (Chapter 5) were determined using a Dionex ICS-2000 ion chromatograph and a Dionex ICS-500 fitted with an IonPac CS12 cation exchange column. Laboratory testing showed detection limits for most major ions to be around 10 µg l^{-1}, although acetate was detectable reliably to 2 µg l^{-1}. Concentrations of HCO_3^- were estimated from charge balance deficits (Hodson *et al.*, 2000). Dissolved Si was determined colourmetrically by the molybdosilic acid method (APHA, 2002) using a Helios Gamma spectrophotometer (Thermo Fisher Scientific).

Accurate measurements of extremely dilute nutrient concentrations were required for research in Chapter 6. NO_3^- and NH_4^+ were determined colourmetrically using an AutoAnalyzer 3 (Bran + Luebbe) with a 1 µg l^{-1} detection limit. PO_4^- was determined by the ascorbic acid method (APHA, 2002) with a detection limit of 5 µg l^{-1}. Analytical precision was typically <5% for all determinands, except NO_3^- when measured by Dionex where precision errors up to 20% were recorded because of very low solute concentrations.

3.7.5 Isotopic analysis

Hydrogen and oxygen stable isotope analyses were conducted using a GV Instruments Isoprime isotope ratio mass spectrometer connected to a Eurovector Environmental Analyser with an analytical precision of <0.9 ‰ for both δ^2H and $\delta^{18}O$. Stable isotope values are expressed using the δ convention in per mil (‰) units relative to Vienna Standard Mean Ocean Water (V-SMOW) as:

$$\delta^2H_{sample}\ (‰) = (^2H / ^1H_{sample}) / (^2H / ^1H_{standard}) - 1$$

and

$$\delta^{18}O_{sample}\ (‰) = (^{18}O / ^{16}O_{sample}) / (^{18}O / ^{16}O_{standard}) - 1$$

3.7.6 Chlorophyll a

Chlorophyll *a* (mg m^{-2}) was determined using a Helios Gamma spectrophotometer (Thermo Fisher Scientific) following extraction for 12 h in 90% acetone and acidification to correct for pheophytin content (Steinman *et al.*, 2006). Cobble areas, required to calculate areal concentrations of chlorophyll *a*, were quantified from digital photographs using ImageJ software (Rasband, 2012).

3.7.7 Macroinvertebrate sorting and identification

The majority of collected taxa were Chironomidae larvae which were identified to species level where possible. Other taxa were identified to family level. Random subsampling was used where abundance exceeded 60 individuals per sample (Lods-Crozet *et al.*, 2001). Chironomids were prepared in 10% KOH solution at room temperature for approximately 12 h, washed in distilled water and slide mounted in dimethyl hydantoin formaldehyde resin. Individuals were identified at x100-600 magnification using various keys (Rossaro, 1980; Wiederholm, 1983; Oliver and Roussel, 1983; Stur and Ekrem, 2011).

3.8 Chapter summary

The research design of the thesis and information on the study area, data sources and sampling methodologies used in later research chapters has been presented. Data processing techniques have also been discussed. These data allow for the identification of conceptual water stores that contribute to basin runoff (Chapter 4), analysis and interpretation of water temperature dynamics (Chapter 5), estimation of nutrient uptake parameters (Chapter 6), and assessment of benthic macroinvertebrate community responses to water source dynamics and physicochemical habitat conditions (Chapter 7) in Arctic rivers.

CHAPTER 4

WATER SOURCING AND HYDROLOGY OF ARCTIC RIVER BASINS

4.1 Introduction

Recent observations show that Arctic cryospheric systems are sensitive to climate variability (Zemp *et al.*, 2009; Walsh *et al.*, 2011). Projected climate change in high-latitude regions is expected to impact strongly on glacier mass-balance, snow distribution and permafrost extent (Anisimov *et al.*, 2007; Callaghan *et al.*, 2011b). Future changes in cryospheric systems are likely to affect high-latitude hydrological processes because Arctic river flow regimes are controlled to a large extent by the release of water from frozen basin stores (Pohl *et al.*, 2007). In turn, these changes in river flows could alter fluxes of freshwater, sediment and nutrients with implications for biodiversity in aquatic and marine ecosystems (Prowse *et al.*, 2006; Anisimov *et al.*, 2007; Milner *et al.*, 2009).

A review of the literature in Chapter 2 showed that there is currently limited understanding of the hydrological processes and key water sources that govern river flow dynamics in Arctic river basins. Hydrochemical and isotopic characteristics of Arctic river waters may allow for the identification of conceptual basin water sources and offer potential for generating quantitative estimates of dynamic water source contributions that contribute to river basin runoff. Such knowledge is necessary to underpin accurate predictions of future flow regimes of high-latitude rivers. Accordingly, the aim of this chapter is to quantify variability in water sourcing in river basins spanning a range of glacierization in Svalbard to yield new process-based understanding of basin-wide hydrological fluxes and dynamics. The specific objectives of the chapter are to:

I. determine the major processes governing river hydrochemistry and isotopic composition;
II. assess the potential of hydrochemical and isotopic signatures to identify distinct conceptual water sources (e.g. snowmelt, ice melt, shallow groundwater, rainfall-runoff), and;
III. understand seasonal and year-to-year variability in water source contributions to river flow.

The implications of the chapter‘s findings are considered in the context of future environmental change and used to infer potential hydrological changes that may occur under projected climatic changes in high-latitude regions.

4.2 Methodology

4.2.1 Study area and sampling framework

The study was undertaken near Ny Ålesund on the Brøgger peninsula from 6 July to 31 August 2010 (calendar days 187-242) and from 4 July to 3 September 2011 (calendar days 185-246). Field investigations were focused on two primary study areas, the Bayelva basin and the Stuphallet cliffs area (Figure 4.1). Field sampling was designed to investigate rivers sourced from a variety of basin water sources including glaciers, snowpacks and groundwater at seven study sites. Three sites in the Bayelva basin were selected to provide a range of glacial influence over which to evaluate water sourcing near the glacier snout (V1), along the river mainstem (V2), and at the Bayelva river mouth (BR). Additional sites in Stuphallet cliffs area (S3; 2010 only) and the wider Bayelva area (N1, N2, N3; 2011 only) were identified to compare between glacierized and non-glacierized systems (Figure 4.1; Table 4.1). The Stuphallet cliffs area was accessed primarily by boat. In 2011, a lack of boat transport meant that sampling efforts were directed closer to Ny Ålesund. Non-glacierized sites were hypothesised to be representative of snowmelt and hillslope groundwater dominated river systems found in this region of Svalbard.

Hydroclimatological variables were recorded continuously at river sites during the 2010 and 2011 meltwater seasons. To understand the major controls governing flows in these Arctic river systems, these data were combined with data from hydrochemical and isotopic samples acquired weekly from river sites and rainfall collectors, and approximately every two weeks from snowpacks, glacier ice and groundwater seeps.

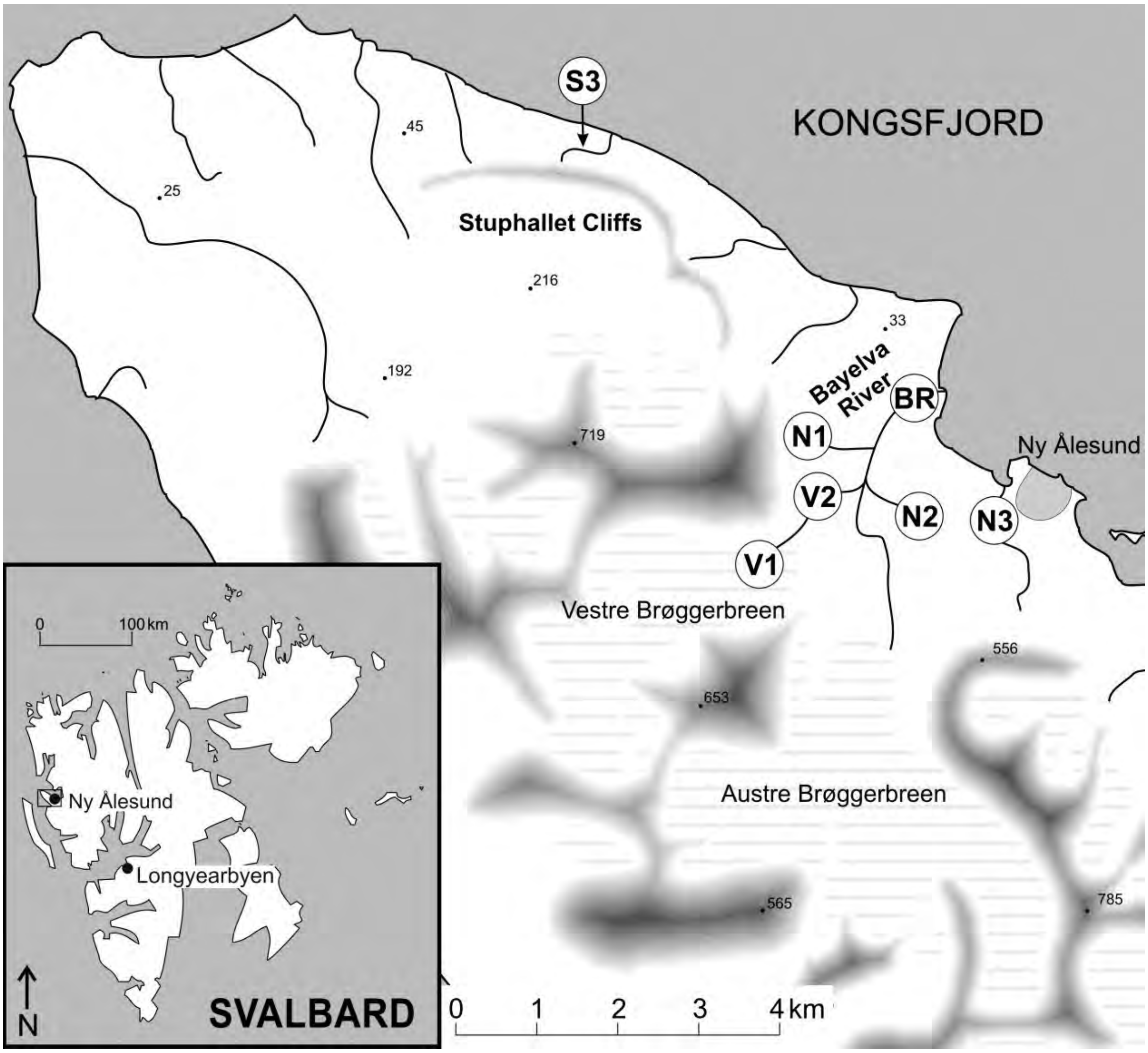


Figure 4.1: *Map of study area showing sampling sites, approximate river courses, relief (shaded), spot height measurements in meters asl, and glacier cover (dashed areas)*

Table 4.1: *Descriptions and major characteristics of the research sites*

Site	*Location*	*Primary water source*	*Distance from glacier (m)*	*Catchment area (km^{-2})*	*Glaciation (%)*
V1	Bayelva	Vestre Brøggerbreen glacier	100	2.1	61
V2	Bayelva	Vestre Brøggerbreen glacier	1000	10.2	49
BR	Bayelva	Austre and Vestre Brøggerbreen glaciers	2500	33.5	54
N1	Bayelva	Snowmelt and groundwater	n/a	0.9	0
N2	Bayelva	Snowmelt and groundwater	n/a	0.4	0
N3	Bayelva	Snowmelt and groundwater	n/a	1.8	0
S3	Stuphallet	Snowmelt and groundwater	n/a	0.6	0

4.2.2 Hydroclimatological monitoring

River discharge was monitored continuously at 15 min resolution at sites V1, V2, BR, and S3 in 2010 and at V1, V2, BR, N1, N2 and N3 in 2011. In 2010, logistical constraints prevented installation of the datalogger at site S3 until day 195. Electrical conductivity (EC) was monitored continuously at 15 min resolution at sites V1 and V2 and BR (2011 only). Air temperature data at 15 min resolution and daily precipitation data were obtained from a weather station in Ny Ålesund approximately 1 km from the Bayelva basin.

4.2.3 Hydrochemical sampling and analysis

River water samples for hydrochemical analysis were collected weekly by hand between DOY 187-242 (2010) and DOY 185-246 (2011) at each site at midday (1100 – 1400 h) to minimise sub-daily patterns influencing longer-term trends in hydrochemical and isotopic variability (Brown, 2002; Cable *et al.*, 2011). Logistical constraints made it impractical to sample both high and low flows. Groundwater samples were collected from hillslope seeps in the Bayelva and Stuphallet cliffs areas every two weeks throughout the study periods. Snow samples were collected from around the Bayelva area. Glacial ice samples were acquired from the lower sections of Austre Brøggerbreen and Vestre Brøggerbreen glaciers.

Major ion and dissolved Si concentrations were determined for all water samples. Crustal solute concentrations (hereafter denoted by prefix _*‘) were derived by using standard seawater ratios to Cl^- to remove the marine-derived fraction of major ions from the dataset (Holland, 1978; Sharp, 1995). Silicate and dolomite weathering rates were examined using *K:Si and $*Ca^{2+}$:$*Mg^{2+}$ ratios, respectively (Hodson *et al.*, 2000; Wadham *et al.*, 2001). Relative rates of carbonate dissolution versus silicate weathering were assessed using $[*Na^+ + *K^+]:[*Ca^{2+} + *Mg^{2+}]$ ratios (Brown *et al.*, 2006b).

4.2.4 Isotopic sampling and analysis

Samples of river water, snow, ice and groundwater were collected for stable isotope analysis in conjunction with hydrochemical sampling. In addition, rainfall samples were collected approximately weekly from site BR, with more frequent samples acquired during heavy rainfall events. Stable isotope analyses were conducted for all samples to determine δ^2H and δ^{18}O values.

4.2.5 Hydrograph separation

Seasonal and year-to-year patterns in water source dynamics at each site were quantified to assess controls on basin-scale hydrological functioning. Hydrochemical and isotopic data were screened to identify the key conceptual water stores which contributed to bulk river flows. These were used subsequently in end-member mixing analyses to separate river hydrographs into individual source components (Christopherson and Hooper, 1992). This method has been used successfully in lower-latitude rivers (McNamara *et al.*, 1997; Liu *et al.*, 2004; Brown *et al.*, 2006b; Dzikowski and Jobard, 2011) but has not previously been applied to glacierized basins in Arctic regions. A steady state form of mass balance equations was used to separate river discharge into two flow components on the basis of end member concentrations (Caine, 1989; Sueker *et al.*, 2000):

$$Q_s = Q_m + Q_g$$

$$C_s Q_s = C_m Q_m + C_g Q_g$$

where *Q* is discharge, *C* is solute or isotopic concentration, and the subscripts *s*, *m* and *g* refer to the river, first end member and second end member components, respectively. End-members were tested for distinctiveness using independent sample *t*-tests.

The uncertainty related to each flow component was quantified using the method outlined by Genereux (1998) to provide confidence limits for each mixing fraction:

$$W_{fm} = \sqrt{\left[\frac{C_g - C_s}{(C_g - C_m)^2} W_{Cm}\right]^2 + \left[\frac{C_s - C_m}{(C_g - C_m)^2} W_{Cg}\right]^2 + \left[\frac{-1}{C_g - C_m} W_{Cs}\right]^2}$$

where *W* is the uncertainty and *f* is the mixing fraction. For end members, the standard deviation of mean solute concentrations were multiplied by appropriate *t* values to give uncertainty estimates at a 95% confidence level. Separate mixing calculations were conducted for each river sample (i.e. no replicates). Therefore, uncertainty for each river component was taken as the analytical precision (Genereux, 1998).

4.3 Results

Seasonal hydroclimatological data are presented to contextualise conditions over the 2010 and 2011 summer melt seasons. Hydrochemical and isotopic characteristics of conceptual water sources and river discharge were utilised to characterise end-member composition for hydrograph separation. Results from mixing analyses were used to show seasonal and year-to-year variability in basin-wide water source dynamics.

4.3.1 Seasonal hydroclimatological context

The 2010 melt season was characterised by cool and dry hydroclimatological conditions (Figure 4.2a). Air temperature ranged from -1.5 ºC to 10.9 ºC with the warmest period in early July (Table 4.2). Precipitation inputs were very low, totalling 17 mm over the study period. Snow cover in the proglacial area of the Bayelva basin was limited to isolated patches at low altitudes in early summer, which retreated rapidly so that ice on the Austre Brøggerbreen and Vestre Brøggerbreen glaciers was exposed by day 195 (14 July) at ~100 m a.s.l.. In 2011, mean air temperature during the melt season was 1.2 ºC higher than 2010 (Figure 4.2b; Table 4.2). Precipitation was markedly higher (totalling 89 mm) than 2010, particularly in early September when total daily precipitation exceeded 16 mm on both days 244 and 245.

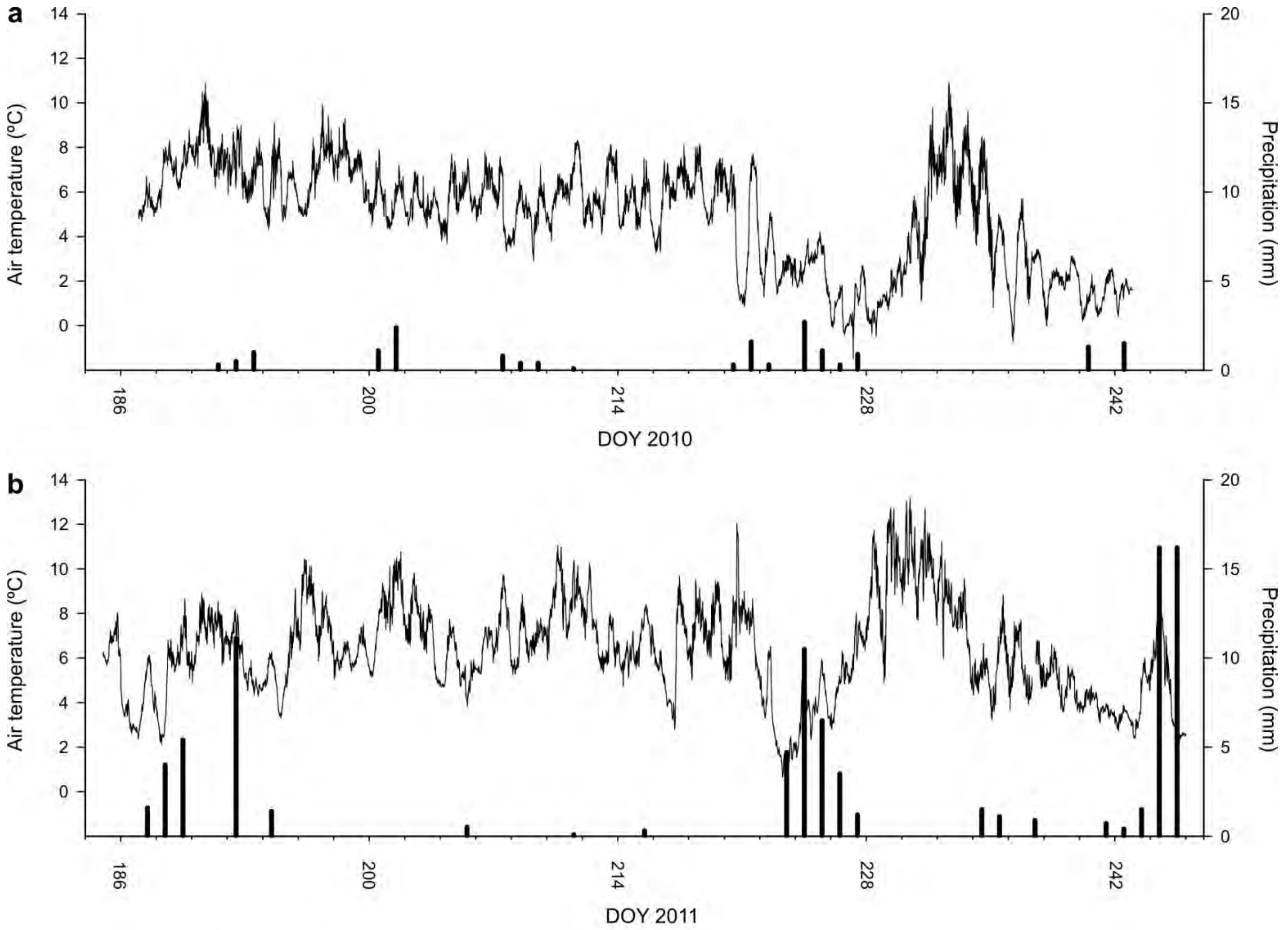


Figure 4.2: *Air temperature and precipitation inputs for a) the 2010 and b) the 2011 melt seasons*

Table 4.2: *Descriptive statistics for air temperature, precipitation, discharge (Q) and specific discharge during the 2010 and 2011 melt seasons*

	Air temperature 2010	*Air temperature 2011*	*Precipitation 2010*	*Precipitation 2011*	*Q 2010*				*Q 2011*					
					V1	*V2*	*BR*	*S3*	*V1*	*V2*	*BR*	*N1*	*N2*	*N3*
	ºC	*ºC*	*mm day^{-1}*	*mm day^{-1}*	*m^3 s^{-1}*				*m^3 s^{-1}*					
Mean (sum)	4.93	6.17	(17.00)	(89.00)	0.89	1.41	2.69	0.04	1.10	2.12	4.19	0.05	0.01	0.08
Max	10.90	13.19	2.70	16.20	3.94	4.99	8.44	0.06	4.72	8.31	11.51	0.07	0.02	0.22
Min	-1.50	0.53	0.00	0.00	0.00	0.01	0.01	0.03	0.13	0.67	0.87	0.03	0.00	0.04
Range	12.40	12.83	2.70	16.20	3.94	4.98	8.43	0.03	4.59	7.64	10.64	0.04	0.02	0.18
Std dev	2.32	2.12	0.61	4.92	0.64	1.07	2.19	0.00	0.52	0.98	1.91	0.00	0.00	0.03

	Specific Q 2010				*Specific Q 2011*					
	V1	*V2*	*BR*	*S3*	*V1*	*V2*	*BR*	*N1*	*N2*	*N3*
	mm day^{-1}				*mm day-1*					
Mean (sum)	37.5	11.9	6.9	5.0	46.4	17.9	10.8	4.4	3.8	1.5
Max	78.2	27.6	16.7	7.2	112.3	48.1	25.5	6.2	8.4	3.6
Min	0.2	0.1	0.0	4.1	9.4	6.0	2.7	3.3	1.9	0.3
Range	78.0	27.5	16.7	3.1	102.9	42.1	22.8	2.9	6.5	3.3
Std dev	24.7	8.5	5.2	0.9	17.4	7.3	4.1	0.7	1.5	1.0

River discharge at glacier-fed sites in the Bayelva basin increased longitudinally downstream with seasonal means in 2010 of 0.89 $m^3 s^{-1}$, 1.41 $m^3 s^{-1}$, and 2.69 $m^3 s^{-1}$ at sites V1, V2 and BR, respectively. In 2011, mean flows at these three sites were 1.10 $m^3 s^{-1}$, 2.12 $m^3 s^{-1}$ and 4.19 $m^3 s^{-1}$, representing increases of 24 %, 51 % and 56 %, respectively (Table 4.2; Figure 4.3). Discharge at all glacier-fed sites exhibited strong diurnal variations, which were related closely to changes in atmospheric energy exchange. In addition, seasonal trends in discharge mirrored those for air temperature dynamics with colder periods towards the end of the melt season corresponding to reduced flow volumes. Discharge at glacier-fed sites displayed a negligible response to precipitation inputs, with the exception of one rainfall event at the end of the 2011 melt season (days 244-245) that resulted in increased flow at all sites (Figure 4.2; Figure 4.3). EC at glacier-fed sites exhibited clear diurnal cycles that were inversely related to discharge (2010: r=-0.76 – -0.84, p<0.001; 2011: r=-0.14 - -0.50, p<0.001; Figure 4.3). EC increased typically by around 20 $\mu S\ cm^{-1}$ between sites V1 and V2 in both years, whereas at site BR EC was similar to site V1. In both years, increased EC at all sites in mid-August was associated with periods of cold air temperature and low discharge at glacier-fed sites.

For the Stuphallet cliffs area in 2010, mean seasonal discharge at S3 (0.02 $m^3 s^{-1}$) was 1-2 orders of magnitude lower than at glacier-fed sites in the Bayelva basin, and diurnal-scale patterns were less apparent. In 2011, discharge at non-glacial sites in the Bayelva area (N1, N2 and N3) was also 1-2 orders of magnitude lower than at glacier-fed sites. Discharge at non-glacierized sites generally displayed low diurnal variability with the exception of site N3 in early July 2011 which exhibited a flow regime similar to glacier-fed sites (Figure 4.3). At site S3 in 2010 discharge was highest at the beginning of the study period and declined steadily as the melt season progressed. Similar reductions in discharge were observed in 2011 at sites N1, N2 and N3, although gradual seasonal declines in flow were punctuated by episodic rainfall events (Figure 4.3). In contrast to glacier-fed sites, non-glacierized river flow showed greater sensitivity to precipitation inputs than to changes in air temperature (Figures 4.2 and 4.3). Mean specific discharge from non-glacierized basins (1.5 – 5.0 mm day^{-1}) was lower and exhibited less variability than at glacier-fed sites (6.9-46.4 mm day^{-1}; Table 4.2).

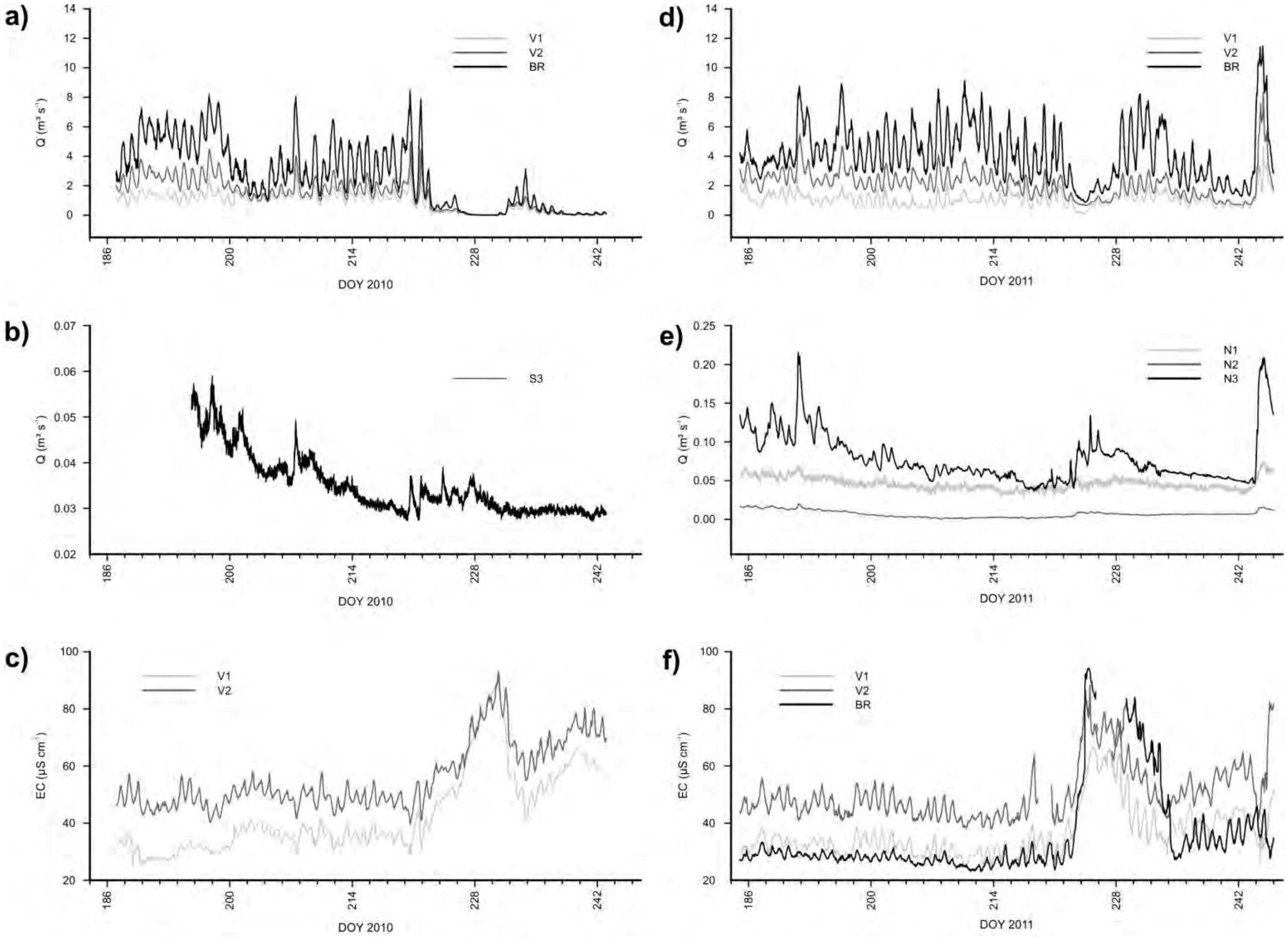


Figure 4.3: *Time series of discharge at sites a) V1, V2 and BR and b) S3 and d) EC at sites V1 and V2 during and 2010 melt season, and discharge at sites d) V1, V2 and BR and e) N1, N2 and N3 and f) EC at sites V1, V2 and BR during the 2011 melt season*

4.3.2 Hydrochemical dynamics of conceptual water sources and river flows

Most water samples were typically enriched in Ca^{2+} with respect to other cations, although snow and ice samples also contained relatively high concentrations of Na^+ (Table 4.3). The dominant anion in all water samples was HCO_3^-. Concentrations of solutes in snow and ice samples were low relative to other water samples, apart from NH_4^+ where the highest concentrations of all samples were found in snow acquired during 2010. Soil waters were enriched in most ions relative to other sample types. Similarly, Si was negligible in snow and ice samples (<0.1 mg l^{-1}) but elevated in soil waters, particularly in the Bayelva basin where mean concentrations reached almost 2 mg l^{-1} in 2010 (Table 4.3).

Solute loads at site V1 near the glacier snout were low relative to other sites and characterised by a dominant Ca^{2+}/HCO_3^- chemistry. Solute concentrations increased downstream in the Bayelva basin for most determinands. Seasonal declines in Cl^- concentration were observed at most sites in both years, although increases were observed in late summer at sites S3 and N2 (Figure 4.4). Mean solute concentrations at non-glacierized sites (S3, N1, N2, N3) were higher than at glacier-fed sites in the Bayelva basin (V1, V2, BR) and ratios of *K:Si were generally lower and less variable (Figure 4.5a). Ratios of $[*Na^+ + *K^+]:[*Ca^{2+} + *Mg^{2+}]$ were low (0.03-0.28) but increased longitudinally away from the glacier and also in non-glacierized basins (Figure 4.5b). Ratios of $*Ca^{2+}:*Mg^{2+}$ were lower at non-glacierized sites in the Bayelva basin, although this pattern did not hold for the non-glacierized site S3 in the Stuphallet cliffs area (Figure 4.5c).

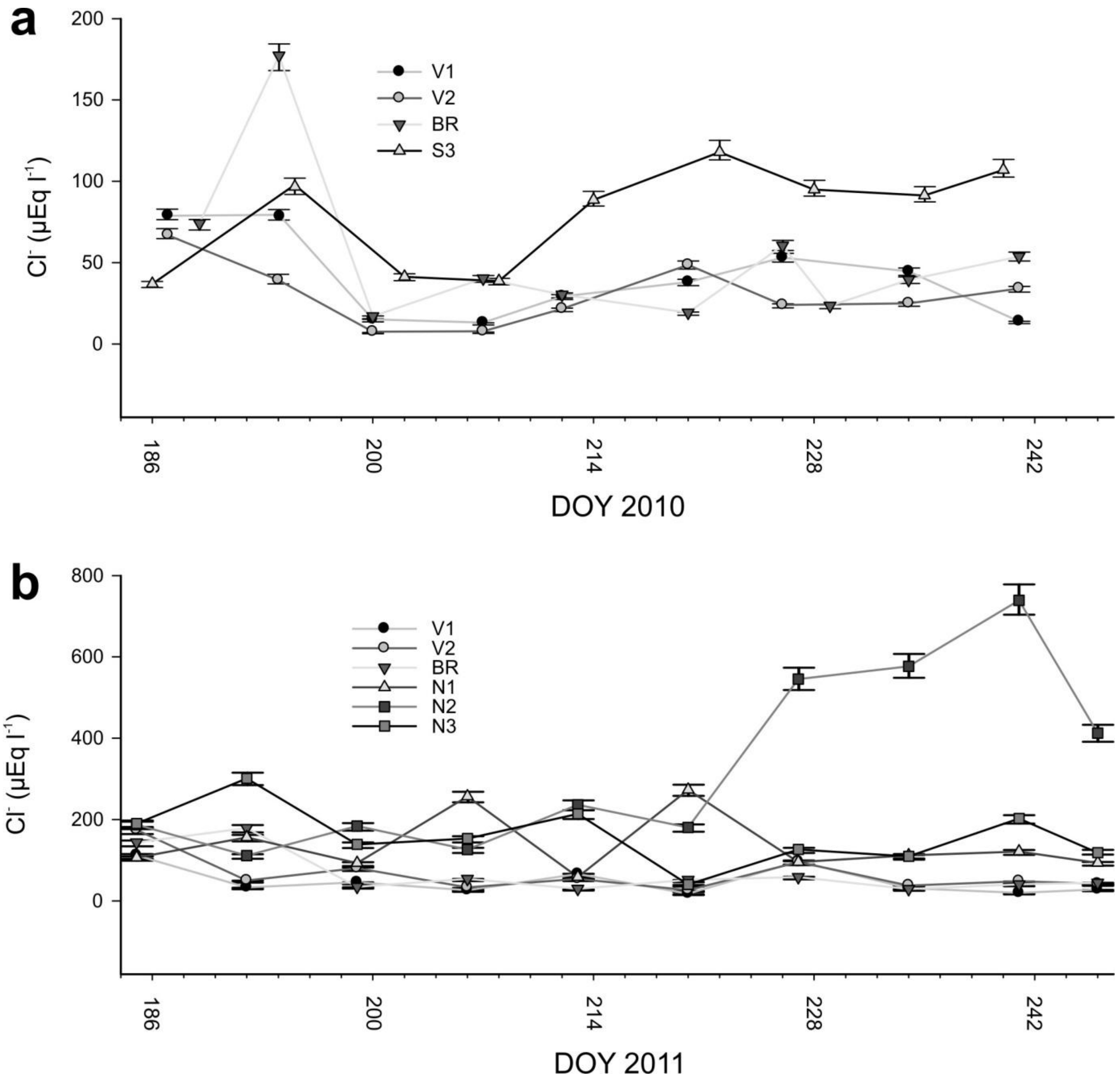


Figure 4.4: *Time series of Cl⁻ concentration for a) the 2010 and b) the 2011 melt seasons. Error bars show 95% confidence limits.*

Table 4.3: *Descriptive statistics for major ions, silica, δ2H and δ18O for water samples acquired during the 2010 and 2011 melt seasons. Mean values and standard deviations (in italics) are provided. *denotes n=11*

Site	*n*	Na^+ ($\mu Eq\ L^{-1}$)		NH_4^+ ($\mu Eq\ L^{-1}$)		Mg^{2+} ($\mu Eq\ L^{-1}$)		K^+ ($\mu Eq\ L^{-1}$)		Ca^{2+} ($\mu Eq\ L^{-1}$)		Cl^- ($\mu Eq\ L^{-1}$)		NO_3^- ($\mu Eq\ L^{-1}$)		SO_4^{2-} ($\mu Eq\ L^{-1}$)		HCO_3^- ($\mu Eq\ L^{-1}$)		Si ($mg\ L^{-1}$)		δ^2H (‰)		$\delta^{18}O$ (‰)	
V1 (2010)	11	32	*13*	3	*2*	27	*12*	34	*14*	490	*78*	32	*17*	1.1	*0.5*	14	*9*	539	*52*	0.3	*0.1*	-78	*5*	-11	*1*
V2 (2010)	9	45	*31*	4	*3*	27	*20*	41	*22*	540	*124*	35	*36*	1.6	*0.4*	15	*10*	606	*59*	0.3	*0.2*	-76	*3*	-11	*0*
BR (2010)	10	49	*42*	5	*4*	45	*26*	52	*39*	516	*179*	46	*43*	1.8	*0.7*	17	*10*	602	*65*	0.3	*0.2*	-80	*4*	-11 *	*1*
V1 (2011)	10	45	*16*	4	*4*	22	*8*	13	*4*	678	*138*	48	*33*	1.4	*0.6*	11	*8*	702	*60*	0.5	*0.1*	-84	*6*	-12	*1*
V2 (2011)	10	52	*9*	4	*3*	28	*10*	13	*3*	767	*154*	64	*44*	1.3	*0.4*	14	*6*	784	*81*	0.5	*0.1*	-83	*5*	-12	*1*
BR (2011)	10	57	*10*	5	*3*	25	*10*	19	*5*	711	*152*	67	*52*	1.7	*1.2*	15	*4*	734	*70*	0.5	*0.1*	-86	*5*	-12	*1*
S3	8	78	*27*	1	*1*	20	*11*	116	*45*	865	*215*	83	*29*	2.5	*1.6*	17	*8*	978	*112*	0.5	*0.3*	-74	*2*	-10	*0*
N1	10	125	*19*	3	*4*	89	*17*	14	*6*	864	*199*	137	*72*	1.9	*1.0*	24	*14*	932	*130*	1.1	*0.3*	-80	*3*	-11	*0*
N2	10	306	*169*	5	*6*	96	*45*	27	*8*	1263	*523*	330	*222*	1.8	*1.6*	30	*20*	1334	*450*	1.0	*0.3*	-81	*4*	-11	*1*
N3	10	234	*92*	5	*3*	120	*44*	32	*11*	1415	*511*	159	*71*	2.4	*1.8*	31	*8*	1612	*725*	1.1	*0.3*	-83	*5*	-12	*1*
Snow (2010)	34	57	*40*	16	*16*	8	*6*	6	*4*	54	*35*	52	*42*	1.7	*0.3*	10	*7*	153	*62*	0.1	*0.0*	-89	*27*	-12	*3*
Ice (2010)	24	29	*18*	5	*4*	4	*3*	4	*3*	28	*20*	38	*27*	0.5	*0.1*	17	*12*	48	*16*	0.1	*0.0*	-81	*6*	-11	*1*
Snow (2011)	15	19	*7*	4	*2*	2	*1*	8	*4*	79	*52*	19	*8*	1.8	*0.5*	6	*4*	84	*43*	0.0	*0.0*	-99	*23*	-13	*2*
Ice (2011)	12	11	*9*	3	*2*	1	*1*	5	*2*	36	*16*	10	*6*	0.4	*0.1*	7	*2*	40	*9*	0.0	*0.0*	-77	*5*	-11	*1*
Groundwater (BR 2010)	29	140	*92*	6	*4*	49	*33*	227	*142*	1118	*694*	116	*145*	4.1	*3.8*	20	*10*	1400	*890*	2.0	*0.7*	-74	*6*	-10	*1*
Groundwater (BR 2011)	54	200	*95*	6	*5*	146	*111*	34	*36*	1363	*848*	282	*210*	3.9	*3.5*	52	*31*	1412	*1054*	1.8	*0.6*	-77	*4*	-10	*1*
Groundwater (S3)	12	139	*86*	11	*7*	45	*27*	163	*95*	1304	*266*	99	*59*	5.2	*4.9*	29	*16*	1529	*834*	1.2	*0.6*	-73	*3*	-10	*0*

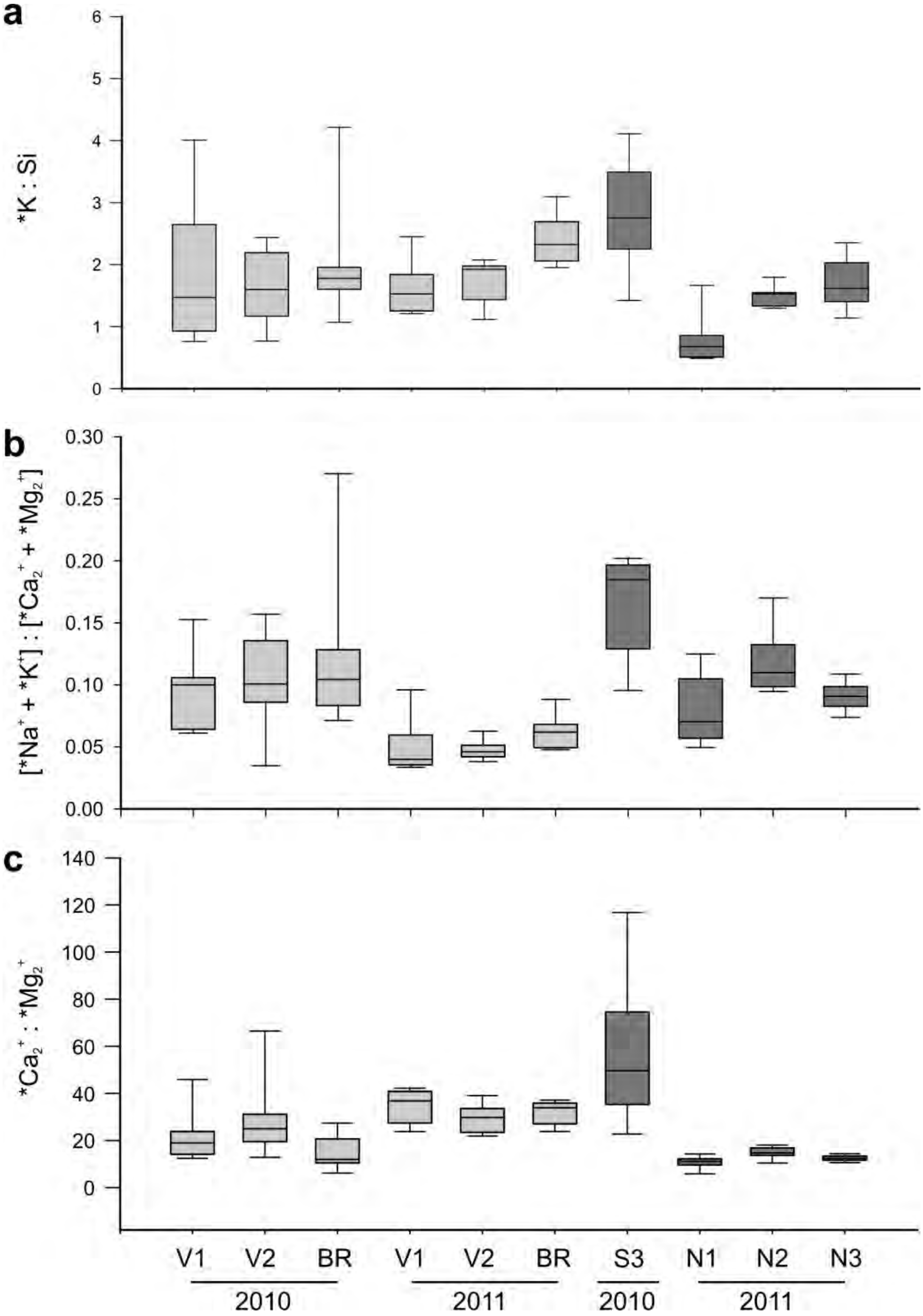


Figure 4.5: *Boxplots of a) *K:Si molar concentrations, b) [*Na$^+$ + *K$^+$]:[*Ca^{2+} + *Mg^{2+}] concentrations, and c) *Ca^{2+}:*Mg^{2+} concentrations by site. n=8-12 for all sites and non-glacierized river sites are shaded dark.*

4.3.3 Isotopic dynamics of conceptual water sources and river flows

The mean isotopic composition across water sample types ranged from -13 and -99 for δ^{18}O and δ^{2}H, respectively, to -10 and -73 for δ^{18}O and δ^{2}H, respectively (Table 4.3). Snow contained the isotopically lightest samples and rain samples were on average most enriched in δ^{18}O and δ^{2}H, although both water types showed high variability in both years. Groundwater seep water samples were slightly enriched in δ^{2}H and δ^{18}O relative to other samples. Variability in the isotopic composition of glacier ice was low and mean values were generally marginally lower than those of river water samples. Samples from glacier-fed sites were typically less enriched in δ^{2}H and δ^{18}O and exhibited more seasonal variability than in non-glacierized basins (Figure 4.6). Trends in δ^{2}H enrichment at site V1 in the Bayelva basin were very similar to sites V2 and BR downstream. Temporal patterns indicated a small increase in δ^{2}H enrichment during the first half of each study period at all sites, followed by a clear decline (>10 δ^{2}H) in late summer for glacier-fed sites only. The late-season decline was much less pronounced (<5 δ^{2}H) at non glacier-fed sites (. In 2011, substantial decreases in river δ^{2}H occurred around days 227 and 246 which coincided with high precipitation (Figure 4.2). These decreases were especially marked at glacier-fed sites.

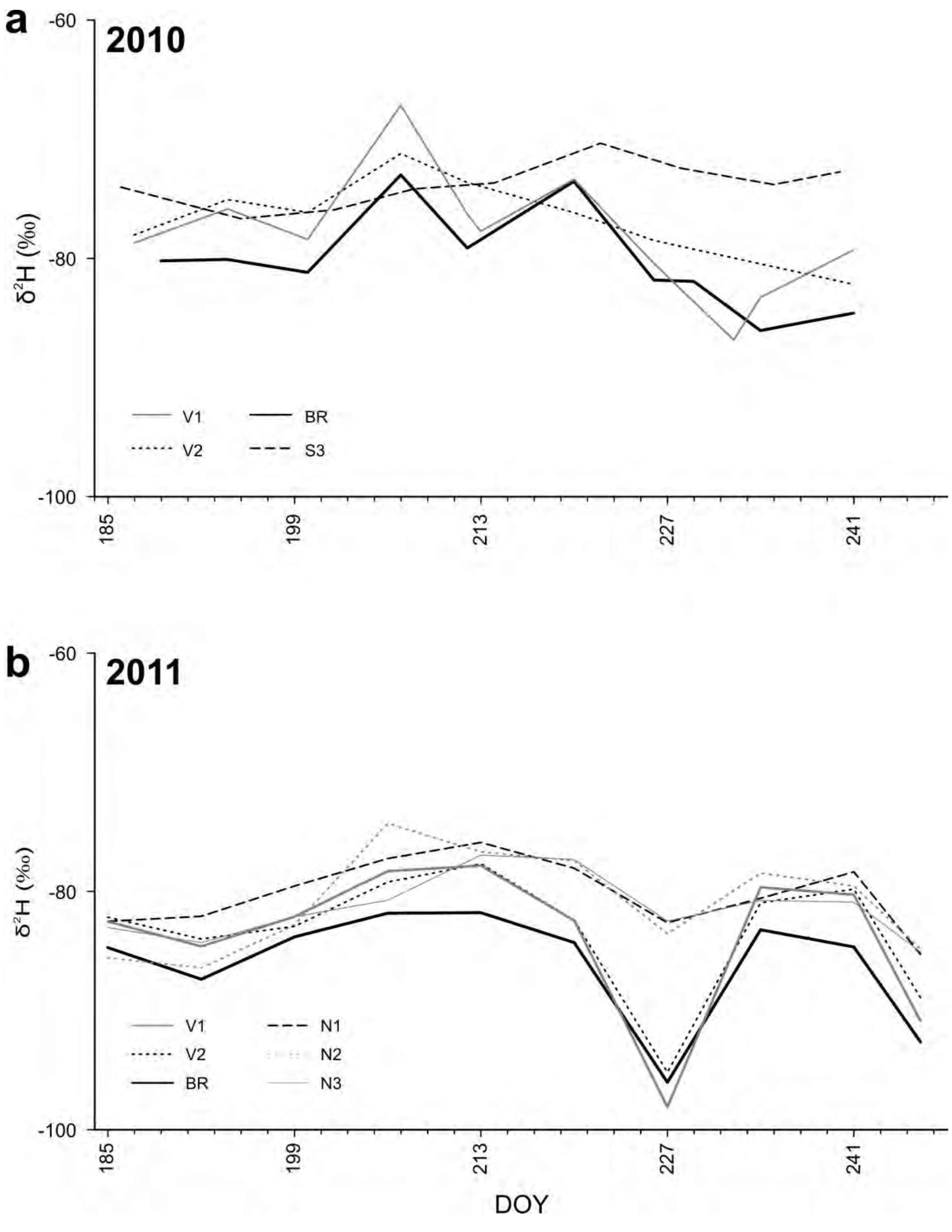


Figure 4.6: *Temporal changes in river water δ^2H isotope composition for sites a) V1, V2, BR and S3 during the 2010 melt season, and b) V1, V2, BR, N1, N2 and N3 during the 2011 melt season.*

4.3.4 End-member characterisation

Screening of hydrochemical and isotopic data showed concentrations of many constituents did not vary significantly between water samples and were thus unsuitable for use in end-member mixing analysis (Christophersen and Hooper, 1992; Hinton *et al.*, 1994). However, independent sample *t*-tests showed that there were significant differences in Si concentration between dilute snow and ice samples and more enriched groundwater seep samples ($p<0.001$ in all sub-basins). These end-members bounded river water samples (Figure 4.7) and made Si a practical tracer of groundwater contributions to bulk river flow (Anderson *et al.*, 2000).

A two-component end-member mixing model was used to separate seasonal hydrographs into individual source components: (i) meltwater, derived from snow and glacial ice ablation (low Si), and (ii) hillslope groundwater, transmitted to the river channel via subsurface flow through the active soil layer (high Si). Although these components could be viewed as ‗pathways' their contrasting chemical signatures enable them to be observed as conceptual water sources (Brown *et al.*, 2006b), thus justifying their use with end-member mixing analysis. The meltwater end-member for sites in the glacierized Bayelva basin (V1, V2, BR) was defined by the mean Si concentration of snow and ice samples for each year, and for non-glacierized sites (N1, N2, N3) by the mean Si concentration of snow samples. As access constraints prevented collection of snow samples in 2010 from the Stuphallet cliffs area, the quickflow end-member for S3 was defined by the mean Si concentration of snow samples from the Bayelva River basin. This was considered appropriate because both spatial variability and concentration of snowpack Si is minimal in precipitation and river load is derived almost entirely from mineral weathering (Campbell *et al.*, 1995). Exploratory analysis revealed no significant patterns or differences in snow or groundwater Si concentration throughout each melt season. Consequently, groundwater components at BR, S3, N1, N2 and N3 were calculated using the mean seasonal Si concentrations of groundwater samples collected in the basin area of each respective sampling site. Steep valley walls and large glacier icepacks limited the number of groundwater samples acquired upstream of sites V1 and V2. Consequently, these sites were assigned the same groundwater component as site BR. Estimated mean uncertainty in the contribution of each component at a 95% confidence level ranged from ±0.08 at site V1 in 2010 to ±0.42 at site N3 in 2011 (Table 4.4). The groundwater component accounted for more than three-quarters of the total uncertainty at each site and was highest at site N2 (99.7 %). The proportion of uncertainty attributable to analytical error in the analysis of river water samples was negligible for all sites (<1%).

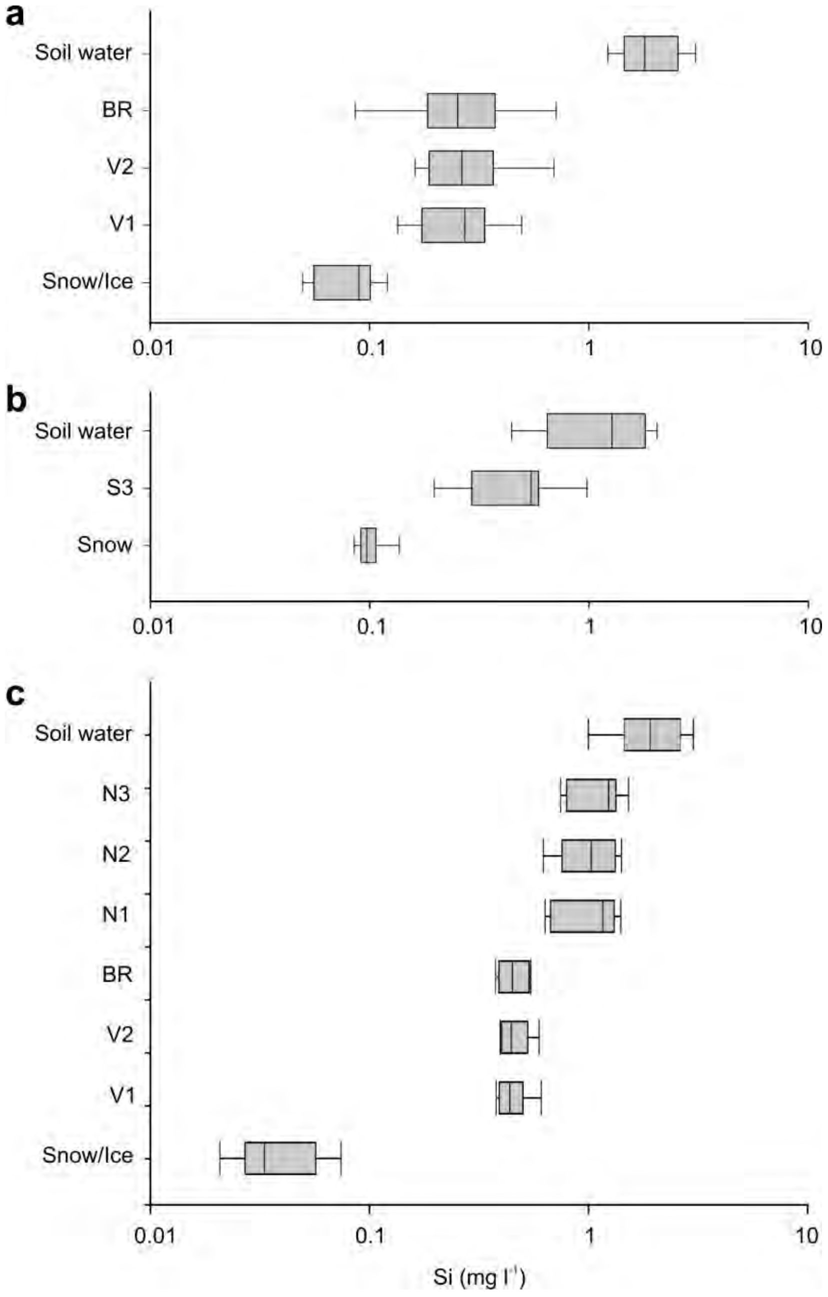


Figure 4.7: *Boxplots of Si concentration for river water samples and meltwater and groundwater end-members for a) sites V1, V2 and BR in 2010, b) site S3 in 2010, and c) sites V1, V2, BR, N1, N2 and N3 in 2011. n.b. Separate end-member concentrations were used for sites N1, N2 and N3, but samples have been combined here for brevity.*

Table 4.4: *Uncertainty estimates (95% confidence) in hydrograph separation using end-member mixing analysis and the percentage of total uncertainty accounted for by each component.*

Site	Uncertainty (95%)			Mean uncertainty accounted for (%):		
	Mean	Max.	Min.	Meltwater	Groundwater	River
V1 (2010)	0.08	0.17	0.03	23.23	75.89	0.88
V2 (2010)	0.09	0.24	0.04	19.06	80.06	0.88
BR (2010)	0.09	0.25	0.03	25.65	73.53	0.83
V1 (2011)	0.15	0.22	0.12	0.04	99.23	0.73
V2 (2011)	0.15	0.21	0.12	0.04	99.24	0.72
BR (2011)	0.15	0.19	0.12	0.04	99.23	0.73
S3	0.42	0.94	0.11	2.37	97.26	0.37
N1	0.36	0.49	0.20	0.00	99.41	0.59
N2	0.36	0.56	0.09	0.01	99.42	0.57
N3	0.42	0.57	0.28	0.00	99.52	0.47

4.3.5 Seasonal and year-to-year variability in water source dynamics

During the 2010 melt season, glacial and snow meltwater dominated flows at glacier-fed sites in the Bayelva basin and accounted for almost 90% of mean river discharge (Table 4.5). The meltwater component was highest (89.7 ± 6.1%) by the snout of the Vestre Brøggerbreen glacier (V1). Both meltwater and groundwater components increased longitudinally downstream, although relative differences in meltwater and groundwater contributions between glacier-fed sites were small (<2%). On a seasonal basis, absolute groundwater contributions at sites in the Bayelva basin in 2010 were greatest in early July, although relative contributions increased towards the end of the study period as total river discharge declined (Figure 4.8). Bulk discharge at S3 was comprised principally of snowmelt (>80%) in mid July but contributions from this source decreased rapidly towards the end of the month and continued to decline throughout the study period so that groundwater contributions became the dominant (>75 %) component of river flow by day 222 (Figure 4.8).

Table 4.5: *Contributions of meltwater and groundwater to river discharge at sampling sites for the 2010 and 2011 melt seasons. Mean values and standard deviations (in parentheses) are provided. Note that mean discharge values derived from EMMA results (n=8-12) differ from those in Table 4.2 calculated from 15 min data (n≈5000).*

Site	$Q_{\text{meltwater}}$ ($m^3\ s^{-1}$)		$Q_{\text{groundwater}}$ ($m^3\ s^{-1}$)		$\%_{\text{meltwater}}$	$\%_{\text{groundwater}}$	
V1 (2010)	1.2	(1.2)	0.1	(0.1)	89.7	10.3	(6.1)
V2 (2010)	1.8	(1.4)	0.2	(0.2)	88.1	11.9	(8.8)
BR (2010)	3.3	(2.9)	0.2	(0.3)	88.2	11.8	(10.9)
V1 (2011)	0.9	(0.35)	0.2	(0.14)	78.9	21.1	(4.7)
V2 (2011)	1.6	(0.66)	0.4	(0.16)	78.5	21.5	(4.3)
BR (2011)	3.0	(1.28)	0.7	(0.27)	79.1	20.9	(4.1)
S3	0.010	(0.006)	0.004	(0.002)	64.6	35.4	(24.2)
N1	0.014	(0.011)	0.034	(0.010)	27.9	72.1	(20.3)
N2	0.002	(0.003)	0.006	(0.004)	28.6	71.4	(25.6)
N3	0.022	(0.023)	0.064	(0.029)	22.5	77.5	(18.4)

In 2011, the mean contribution of groundwater to total river flow at glacier-fed sites (21 %) increased by approximately 10% compared to 2010. However, meltwater remained the dominant flow component throughout the study period at V1, V2 and BR with little longitudinal or seasonal changes in groundwater contributions (Table 4.5; Figure 4.8). At sites in non-glacierized basins (N1, N2, N3), seasonal declines in the meltwater component and progressive increase in groundwater contribution to total river flow were similar to that observed at S3 during 2010. The dominance of soil water contributions to river discharge were such that this fraction comprised up to 100% of total flow at N2 and N3 during August 2011. In contrast to 2010, the meltwater flow component increased slightly at most sites at the end of the 2011 study period (day 246). In both years, seasonal variability in water source contributions was higher at sites in non-glacierized basins (SD: 18.4-25.6 %) compared to glacier-fed sites (4.1-10.9 %).

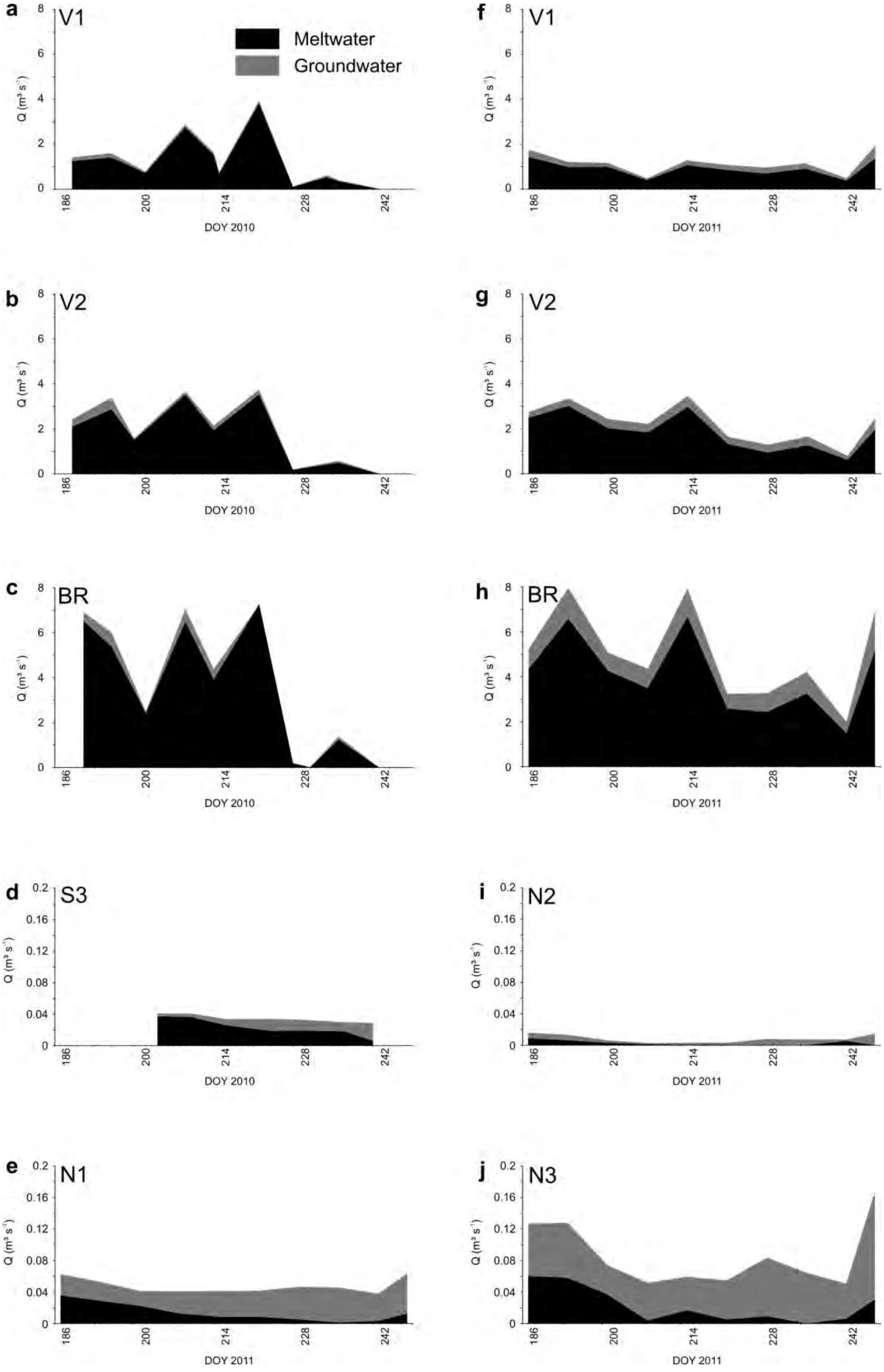


***Figure 4.8:** Seasonal hydrograph separation calculated by end-member mixing analysis for sites a) V1 (2010), b) V2 (2010), c) BR (2010, d) S3 (2010), e) N1 (2011), f) V1 (2011), g) V2 (2011), h) BR (2011), i) N2 (2011), and j) N3 (2011).*

4.4 Discussion

New insights into the basin-scale hydrological processes which govern flow regimes in high-latitude river systems have been provided by characterising variability in water sourcing to rivers in Svalbard during the 2010 and 2011 meltwater seasons. Hydrochemical and isotopic signatures reflected clear spatiotemporal differences in chemical weathering processes, solute provenance and hydrological functioning. By identifying conceptual water stores, quantitative estimates of seasonal and year-to-year dynamics in water source contributions to flow in Arctic rivers have been generated.

4.4.1 Processes controlling basin-scale hydrochemical and isotopic composition

Spatiotemporal variations in hydrochemical and isotopic characteristics were defined clearly and reflected the different hydrological processes occurring in spatially-distinct basin areas over each meltwater season. Several key factors governed river hydrochemical and isotopic signatures during 2010 and 2011. Diurnal cycles in glacier ice and snowpack ablation, driven by atmospheric energy receipt, led to the sub-daily dilution of solute loads in the Bayelva river basin (Hodson *et al.*, 1998b). These cycles were evident in the clear inverse relationship between discharge and electrical conductivity at all glacier-fed sites, suggesting that hydrological processes in upper glacierized basin areas exerted a strong influence on river hydrochemistry in the proglacial zone (Brown *et al.*, 2006b; Dzikowski and Jobard, 2011).

Rapid weathering of exposed glacial moraines in ice-marginal areas led to elevated concentrations of Ca^{2+} and HCO_3^- at site V1 adjacent to the Vestre Brøggerbreen glacier, most likely as a consequence of dilute meltwaters which are especially chemically reactive (Hodgkins *et al.*, 1998). Interestingly, concentrations of SO_4^{2-} at V1 were low relative to HCO_3^-, implying that sulphide oxidation in distributed flow systems below the glacier is minimal due to anoxic conditions at the glacier bed (*cf.*

Lamb *et al.*, 1995; Tranter *et al.*, 1996; Tranter *et al.*, 2002). Previous studies have suggested that Vestre Brøggerbreen has a non-temperate thermal structure (Hodson *et al.*, 2002) but these findings provide the first comprehensive empirical evidence to support this conclusion. Weathering processes at V1 were dominated by carbonate dissolution as indicated by high $*Ca^{2+}:*Mg^{2+}$ ratios, thus reflecting the geology of the upper Bayelva basin (Hjelle, 1993). Similar ratios at other glacier-fed sites demonstrated that ion exchange in meltwaters was limited during downstream transport (*cf.* Wadham *et al.*, 1998), and silicate dissolution was low relative to carbonate weathering as indicated by low ratios of $[*Na^{+} + *K^{+}]:[*Ca^{2+} + *Mg^{2+}]$ (*cf.* Brown *et al.*, 2006b). However, lower $*Ca^{2+}:*Mg^{2+}$ ratios at non-glacierized sites in the Bayelva basin signified increased dolomite weathering in the proglacial area (*cf.* Malard *et al.*, 1999; Wadham *et al.*, 2001).

Seasonal depletion of snowpacks across the Brøgger peninsula resulted in progressive declines in Cl^{-} concentrations at the majority of river sites during both years. For glacier-fed sites, these hydrochemical changes largely reflected the increased dominance of icemelt over snowmelt in the latter half of each study period (Hodson *et al.*, 1998b; Wadham *et al.*, 1998; Rutter *et al.*, 2011). Seasonal declines in discharge in non-glacierized rivers accompanied Cl^{-} reductions at most sites, demonstrating the importance of snowpacks for runoff generation in the absence of glacial ice-melt. However, Cl^{-} increases at S3 and N2 in late-August were most likely due to a concentration effect from high evaporation since there are no documented crustal (halite) sources of Cl^{-} in these basins (Hjelle, 1993; Harland, 1997).

Reduced $*Ca^{2+}:*Mg^{2+}$ ratios at most river sites in non-glacierized river basins (except S3) reflected increased dolomite weathering in lower basin areas and demonstrated the importance of geochemical processes in the active layer for enhancing solute fluxes from these high-latitude river basins, as seen elsewhere in Svalbard (Wadham *et al.*, 2001; Cooper *et al.*, 2002). Similarly, longitudinal increases in EC between V1 and V2 indicated a possible solute source between these sites (Hodson *et al.*, 1998b). The decline in EC further downstream was most likely due to mixing of ionically-dilute meltwaters from Austre Brøggerbreen with more enriched waters from Vestre Brøggerbreen between V2 and BR, as previously inferred by water temperature dynamics (Blaen *et al.*, *in press*). While silicate dissolution was low relative to carbonate weathering, ratios of [$^*Na^+ + {}^*K^+$]:[$^*Ca^{2+} + {}^*Mg^{2+}$] increased in non-glacial river basins. High and variable *K:Si ratios in glacier-fed streams indicated that silicate weathering was largely nonstoichiometric in these systems (Hodson *et al.*, 2000), but increased in importance in most non-glacierized river basins. This finding was a consequence probably of increased rock-water contact and longer residence times in proglacial sediments (Anderson *et al.*, 2000) which, combined with higher water temperature associated with these systems, is known to promote silicate weathering (Anderson, 2005; Blaen *et al.*, *in press*). Therefore, Si was deemed to be a useful tracer of groundwater flow in rivers for this region of Svalbard.

Relatively enriched δ^2H signals at sites in non-glacierized river basins signified high rates of evaporation in the proglacial zone (Cooper *et al.*, 2011). The mean evaporation rate for ice-free areas in Svalbard is around 80 mm a^{-1} (Killingtveit *et al.*, 2003). Potential evaporation in summer is driven largely by high incoming solar radiation during the polar day, and evaporative water losses are believed to represent the greatest output from the proglacial hydrological system (Cooper *et al.*, 2002). In this study, river water samples, particularly those from non-glacierized river basins, were most enriched with δ^2H during mid July during periods of warm air temperature with no precipitation. More negative δ^2H values observed later in each melt season were likely due to reduced atmospheric energy receipt during late August, as reflected by lower air temperature (Hodson *et al.*, 1998b; Blaen *et al.*, *in press*). Additionally, the seasonal development of active soil layers creates deeper water storage potential, promoting infiltration and subsurface storage and thus decreasing evaporation (Cooper *et al.*, 2002; Merck *et al.*, 2012).

The observed variations in isotopic patterns between glacier-fed sites and those in non-glacierized river basins were hypothesised to be due to differences in the routing of water to river channels from different basin stores. During periods of cold weather and increased precipitation, glacier ice ablation was inhibited and evaporative losses lower. Total discharge at glacier-fed river sites dropped and became increasingly dominated by rainfall-runoff with a low δ^2H signal. In contrast, isotopic patterns in water samples from non-glacier fed sites exhibited less sensitivity to climatic variations, most likely because a large proportion of river flow reached the channel via subsurface flowpaths. Seasonal active soil layers in proglacial areas released water gradually to rivers, thus diurnal-scale fluctuations in discharge were less apparent than at glacier-fed sites. Precipitation can infiltrate soil layers and displace existing waters into river channels by shallow subsurface throughflow (Lischeid *et al.*, 2002; Cooper *et al.*, 2011). Consequently, although discharge in non-glacierized river basins responded rapidly to precipitation events, these flows were believed to represent ‗older' displaced water rather than direct rainfall-runoff (*cf.* Sueker *et al.*, 2000; Yang *et al.*, 2012). Hence, the isotopic signal did not vary to the same degree as sites in glacier-fed rivers. One exception to this pattern occurred in late August 2011 when heavy precipitation is thought to have exceeded the infiltration capacity of the active soil layer and led to overland flow which increased river discharge and depleted δ^2H values at all sites (Choiński, 1989; Jones and Rinehart, 2010).

4.4.2 Hydrochemical and isotopic signatures of conceptual water sources

Two conceptual water sources were identified by end-member mixing analysis: meltwater and shallow groundwater. The assumption of two primary water sources was deemed appropriate because Austre Brøggerbreen and Vestre Brøggerbreen are believed to be predominantly cold-based glaciers (Hodson *et al.*, 2002) which lack the distributed subglacial flow components that typify temperate ice structures (Tranter *et al.*, 1993; Brown *et al.*, 2006b). This assumption was supported by the low SO_4^{2-} concentrations in Vestre Brøggerbreen meltwaters. Furthermore, deep groundwater inputs on the Brøgger peninsula were assumed to be negligible because vertical connectivity to deep groundwater reservoirs is limited by permafrost lenses in high-latitude regions (Carey and Woo, 2001; Judd and Kling, 2002; Haldorsen *et al.*, 2010; van der Ploeg *et al.*, 2012). Precipitation inputs were low during 2010 and most of 2011. Therefore, rainfall-runoff was not considered to exert a strong influence on river flow dynamics during most of the study periods. Dissolved Si was identified as a useful tracer of water flow through the active soil layer because relatively high concentrations in groundwater samples contrasted markedly with those of dilute snow and glacier ice. Moreover, the short time (typically < 1 day) required for Si dissolution in the soil matrix justifies the use of this constituent as a quasiconservative tracer of subsurface flow in glacial environments (Malard *et al.*, 1999; Scanlon *et al.*, 2001; Hodson *et al.*, 2002). Unlike previous research in alpine basins (Cable *et al.*, 2011), variability in δ^{2}H and δ^{18}O within and between end-members was high, therefore isotopic data were not considered to be useful in mixing analyses. Some earlier studies utilising chemical mixing models have used variable solute concentrations to reflect temporal variation in end-member composition (e.g. Brown *et al.*, 2006b). However, in this study there were no discernable trends in the composition of meltwater or groundwater end-members, and thus annual end-member concentrations were fixed.

4.4.3 Seasonal variability in water source contributions to river flow

End-member mixing showed clear seasonal variations in water source dynamics derived from the mixing of snow- and glacier-melt with shallow active layer groundwater. Uncertainty in water source estimates was similar to previous studies in glacierized basins (e.g. Brown *et al.*, 2006b) and mainly attributable to the wider Si concentration range spanned by groundwater samples relative to extremely dilute snow samples. In glacier-fed rivers, meltwater dominated flows at all sites, with the highest contributions observed at the beginning of each study period when snow cover was most prevalent and the active soil layer was supposed to be thinnest, as observed in other Arctic regions (McNamara *et al.*, 1997). Interestingly, there was little longitudinal variation in groundwater contribution to bulk river flow at sites in the Bayelva basin, suggesting a limited active layer reservoir compared to glacier runoff. This contrasts with other studies that have adopted similar mixing model approaches in glacierized river systems (e.g. Brown *et al.*, 2006b), where the relative contribution to bulk flow from non-glacial sources increased rapidly with distance from the glacier snout. Hodson *et al.* (2002) suggested leaching of water from soil zones enhances major ion concentrations in proglacial rivers in the Bayelva basin. Our findings support the premise that solute loads increase downstream due largely to high ionic concentrations in groundwater, as seen elsewhere in Svalbard (Cooper *et al.*, 2002; Dragon and Marciniak, 2010), rather than a consequence of substantial volumes of water moving from the active layer to river channels. The relative paucity of groundwater input, particularly during 2010, is likely due to minimal soil development in much of the proglacial area which limits water storage capacity in this comparatively young landscape (Choiński, 1989; Liu *et al.*, 2004), combined with high rates of evaporation as indicated by our isotopic data (Cooper *et al.*, 2002; Lee *et al.*, 2010; Yang *et al.*, 2012). This hypothesis is strengthened by data from non-glacierized river basins where groundwater contributions to bulk flow were very high but specific runoff levels low, indicating ablation in the upper areas of glacierized basins played an important role in maintaining high flow volumes throughout both study periods in glacier-fed rivers, as observed previously in other Arctic and sub-Arctic regions (Marsh and Woo, 1981; Barnett *et al.*, 2005; Dahlke *et al.*, 2012).

Declines in meltwater contributions to river flow in non-glacierized basins during 2010 and 2011 reflected the loss of seasonal snowpacks throughout the melt seasons and were supported by concurrent declines in Cl^- concentrations (Rutter *et al.*, 2011). Increases in groundwater discharge over the same time period were likely due to an increase in soil hydrological storage capacity as the thickness of the active layer expanded throughout the melt seasons (Hinzman *et al.*, 1991; McNamara *et al.*, 1997; Wadham *et al.*, 2001; Townsend-Small *et al.*, 2011). River hydrochemistry reflected this through increased concentrations of most major ions associated with chemical weathering in the developing active layer (Hodson *et al.*, 2002; Stutter and Billet, 2003). Moreover, higher NH_4^+ concentrations in groundwater samples indicated that active layer soils may function as a nutrient source to some Arctic rivers (Rowland *et al.*, 2010). The reduced diurnal fluctuations in discharge in non-glacierized rivers than at glacier-fed sites in the Bayelva basin, especially in late summer of both years, were due to increasingly stable flow contributions from groundwater during these periods (Malard *et al.*, 1999; Brown *et al.*, 2003). Therefore, groundwater contributions appear to play an important role in sustaining flows in non-glacierized rivers in this region of Svalbard following the cessation of annual snowpack melting.

4.4.4 Year-to-year variability in water source contributions to river flow

The relatively short timeframe of this study complicates attempts to assess long-term changes in basin hydrology in this region of Svalbard. However, historic sampling records from NVE at site BR show mean July/August discharge during the period 1990-2007 was 4.5±0.8 $m^3 s^{-1}$, and therefore the two-year dataset of this study can be considered representative of ‗typical' melt seasons. Year-to-year variability in river flow regimes were linked to basin water source dynamics. V1, V2 and BR were meltwater-dominated in both 2010 and 2011; however, higher groundwater contributions during 2011 at all sites may have reflected warmer air temperature during this year, leading to increased active layer thickness and enhanced silicate weathering (Roth and Boike, 2001; Anderson, 2005). Water source dynamics in non-glacierized basins followed a similar trend in both 2010 and 2011, where an initial pulse of meltwater was followed by a gradual increase in groundwater contribution to river flow. These patterns are analogous to those observed in some alpine basins where groundwater reservoirs are recharged progressively by meltwater inputs in spring (e.g. Malard *et al.*, 1999). However, in Svalbard the depth of water infiltration is likely to be considerably lower than in lower-latitude glacierized environments due to the underlying permafrost layers which restrict shallow groundwater storage and upwelling of deep groundwater in many Arctic basins (Hagen and Lefauconnier, 1995; Haldorsen *et al.*, 2010).

4.5 Conclusions and implications

This research explored variability in water sourcing to rivers in several basins in Svalbard during the 2010 and 2011 meltwater seasons. River hydrochemistry and isotopic characteristics reflected spatially-distinct hydrological processes operating over diurnal to seasonal timescales (Objective I), which were used subsequently as a basis for defining conceptual water storage zones within each basin (Objective II). End-member mixing models were employed to assess seasonal and year-to-year variability in water source contributions to bulk flow regimes (Objective III). These models demonstrated that the majority of river flow in all river basins was supplied by dilute meltwater at the beginning of each melt season when snow cover was most extensive and the active layer was shallowest. Meltwater dominated flow dynamics throughout both summers in glacierized basins, although groundwater inputs increased throughout both 2010 and 2011 with seasonal development of the active layer. These groundwater inputs comprised up to 100% of total flow volume in late August in non-glacierized basins, suggesting groundwater sources played an important role in sustaining flows in these river systems throughout the summer following the depletion of seasonal snowpacks.

The relatively short two-year dataset of this study limits prediction of the precise nature of future flow regimes in these basins. Nevertheless, these results provide a basis on which to consider the potential response of Svalbard river systems to climatic changes during the 21st Century. Increased warming and higher precipitation projected for Arctic regions during the 21st Century (Holland and Bitz, 2003; Førland and Hanssen-Bauer, 2003; Anisminov *et al.*, 2007) are likely to have substantial impacts on water source dynamics in high-latitude river basins. Shifts in the annual hydrograph are expected to occur due to the earlier onset of snowmelt (Milner *et al.*, 2009), expansion of the melt season (Durand *et al.*, 2011) and long-term decline in glacial inputs as ice-masses shrink (Lappegard *et al.*, 2007; Moore *et al.*, 2009). However, some evidence seems to suggest that future reductions in meltwater inputs could be compensated for in part by increased discharge from deep groundwater reservoirs associated with the thermal degradation of permafrost (Walvoord and Streigl, 2007; Adam and Lettenmaier, 2008). In this study, groundwater inputs did not contribute considerably to river flow relative to glacier meltwater runoff, and the specific discharge of rivers in non-glacierized basins was substantially lower than those in glacierized basins. This underlines the importance of glacial runoff in sustaining river flows in the period following snowmelt (Barnett *et al.*, 2005). The role of the active layer is likely to increase with thawing, and isotopic data from this study indicate that future increases in precipitation could lead to episodic flushing of active layer waters. This will have associated implications for the export of nutrients from river basins given the large quantity of organic matter stored in high-latitude tundra systems (Tarnocai *et al.*, 2009; Keuper *et al.*, 2012) and the elevated NH_4^+ concentrations observed in groundwater samples.

4.6 Chapter summary

Research in this chapter has examined the key basin-scale processes that underpin seasonal and year-to-year water source dynamics in Svalbard rivers. The results build on previous studies of Arctic basin hydrology and functioning (e.g. Marsh and Woo, 1981) to understand the processes that underpin changing water source dynamics in Arctic river systems both now and potentially under a changing climate. Research in the following chapters considers the implications of such hydrological changes on river habitat properties, ecosystem functioning and aquatic biodiversity in Arctic regions.

CHAPTER 5

WATER TEMPERATURE DYNAMICS IN ARCTIC RIVER BASINS

5.1 Introduction

Forecasted environmental change at high latitudes associated with climatic forcing is likely to affect water temperature dynamics in Arctic rivers (Caissie, 2006; Anisimov *et al.*, 2007; Lammers *et al.*, 2007; Vavrus *et al.*, 2012). Thermal regimes affect ecosystem structure and functioning directly by influencing metabolic rates, physiology and species life-history traits (Poole and Berman, 2001; Füreder, 2007; Friberg *et al.*, 2009). Water temperature also acts indirectly to mediate species interactions (Rahel and Olden, 2008), influence dissolved oxygen levels and the bioavailability of heavy metals (Isaak and Hubert, 2001; Fritioff *et al.*, 2005), and vary transmission rates of parasites and infectious disease (Marcogliese, 2008). River thermal regimes are highly dynamic over multiple spatiotemporal scales (Arscott *et al.*, 2001) and are controlled by numerous factors including atmospheric conditions, topography, discharge and hyporheic exchange (Caissie, 2006; Hannah *et al.*, 2008, 2009).

A literature review in Chapter 2 illustrated that despite the high sensitivity of polar regions to climate change, and the strong influence of temperature upon ecosystem processes, contemporary understanding of water temperature dynamics in Arctic rivers is limited. Greater knowledge of this subject is necessary to develop an accurate assessment of their thermal response to climate change. Consequently, the aim of this chapter is to characterise spatiotemporal water temperature dynamics in Arctic rivers and assess the primary drivers of thermal variability in these systems by testing the following hypotheses:

I. river temperature is determined principally by water source and prevailing hydroclimatological conditions,

II. groundwater inputs lead to relatively warmer and more variable thermal regimes than rivers sourced primarily from glacial runoff, and

III. river water temperature in Svalbard is cooler and more stable compared to alpine rivers at lower latitudes as a consequence of the colder climate and reduced incoming solar radiation in the High Arctic.

The implications of this research are considered in the context of future environmental change and contextualised using current knowledge of water temperature dynamics in Arctic and alpine systems.

5.2 Methodology

5.2.1 Study sites and sampling framework

The study was undertaken from 6 July to 31 August 2010 (calendar days 187 to 242) at eight sites near the Ny Ålesund research station in north-west Spitsbergen (Figure 5.1; Table 5.1). Sites in the Bayelva basin were selected to characterise water column temperatures near the glacier snouts (V1 and A1), along the river mainstem (V2 and A2), and at the Bayelva river mouth (BR). Non-glacial sites in the Stuphallet cliffs area (S1, S2, S3) were considered to be representative of snowmelt and groundwater dominated river systems found in this region of Svalbard.

Water column temperature was measured continuously at each site for the duration of the study period. Additionally, local air temperature was measured at sites V1 and S2. River discharge data was acquired for sites V1, V2, BR and S3. Meteorological data (short-, long- and all-wave incoming radiation (SW↓, LW↓ and AW↓, respectively); air temperature; relative humidity) at 15 min resolution and daily precipitation totals were employed to characterise atmospheric influences on water temperature.

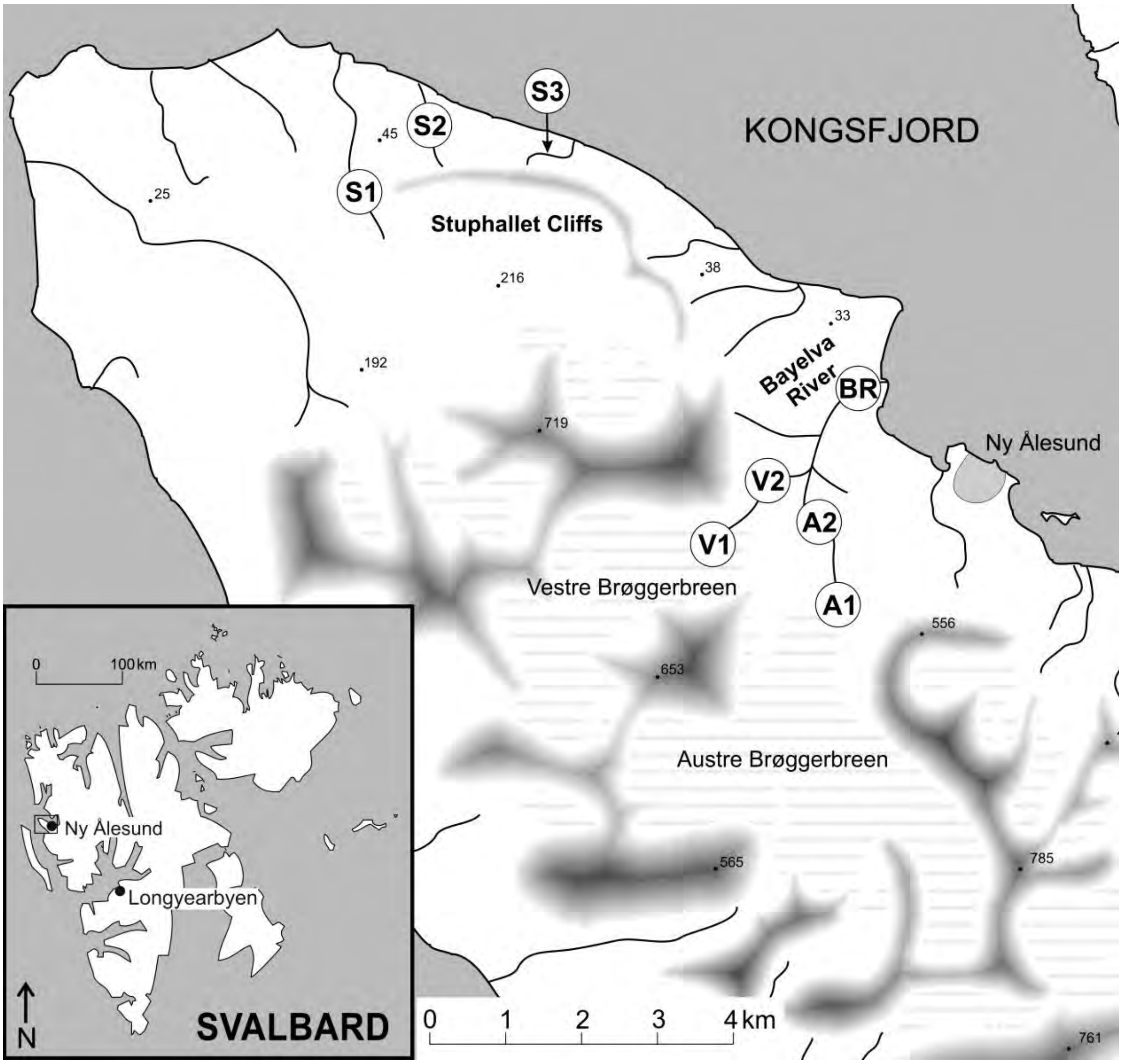


Figure 5.1: *Map of study area showing sampling sites, approximate river courses, relief (shaded), spot height measurements in meters asl, and glacier cover (dashed areas).*

Table 5.1: *Descriptions and major characteristics of sampling sites*

Site	*Location*	*Primary source*	*Distance from glacier (m)*	*Catchment area (km^2)*	*Aspect (facing)*	*Mean EC (μS cm^{-1})*
V1	Bayelva	Vestre Brøggerbreen glacier	100	2.1	E	43
V2	Bayelva	Vestre Brøggerbreen glacier	1000	10.2	E	56
A1	Bayelva	Austre Brøggerbreen glacier	100	9.4	N	-
A2	Bayelva	Austre Brøggerbreen glacier	1000	9.7	N	35
BR	Bayelva	Austre and Vestre Brøggerbreen glaciers	2500	33.5	NE	58
S1	Stuphallet	Snowmelt and groundwater	n/a	4.6	N	-
S2	Stuphallet	Snowmelt and groundwater	n/a	0.9	N	120
S3	Stuphallet	Snowmelt and groundwater	n/a	0.6	N	-

5.2.2 Data analysis

Temperature-duration curves were constructed to compare overall variation in water column temperature at each sampling site (Brown *et al.*, 2005; Hannah *et al.*, 2009). Temperature-duration curves indicate the fraction of time that a temperature was equalled or exceeded at each site during the monitoring period. Average values for each 15 min time-step were computed across the entire melt season and selected 5-day periods to generate composites of temperature variation over the diurnal cycle (Uehlinger *et al.*, 2003). One-way ANOVA was used to evaluate seasonal differences in water temperature variables (mean, max., min., range, std. dev.) between glacial and non-glacial rivers. The degree of association between water-column temperature and hydroclimatological variables (Q, air temperature, short-wave, long-wave and all-wave incoming radiation, relative humidity) was assessed using Pearson‘s product moment correlation coefficients (*r*). Air-water column temperature relationships were calculated using air temperature measured at Ny Ålesund, V1 and S2. Five 5-day sub-periods were selected from the full time-series to characterise diurnal-scale variability in the dataset. These periods were selected to represent the range of hydroclimatological conditions experienced during the summer monitoring period (*cf.* Cadbury *et al.*, 2008) and were set by strong contrasts in air temperature, discharge and precipitation (Hodson *et al.*, 1998b). Pearson‘s product moment correlation coefficients between water temperature and hydroclimatological variables were calculated for each period, and diurnal water temperature composites (described above) were derived to yield visual patterns in water temperature for each period. Seasonal hysteresis patterns between discharge and water temperature were explored using values for each 15 min timestep averaged over the summer monitoring period.

5.3 Results

The results are presented at nested temporal scales. Water temperature patterns were investigated over the 55-day monitoring period; these stream temperature analyses are set in a hydroclimatological context to infer key drivers and processes, using climatological and river discharge data. To examine diurnal patterns in stream temperature in greater detail, five periods of 5 days each were selected that spanned the range of hydroclimatological conditions experienced across the field seasons.

5.3.1 Seasonal hydroclimatological context

The 2010 melt season was characterised mainly by cool, dry hydroclimatological conditions with the warmest air temperatures recorded in early July coinciding with higher AW↓ (Figure 5.2). Cooler air temperature later in the season was associated with decreases in incoming SW↓ under heavily overcast conditions. Air temperature was highest invariably at V1 and lowest in Ny Ålesund (seasonal means 5.4 ºC and 4.9 ºC, respectively), although all air temperature time-series were strongly inter-correlated ($r>0.96$, $p<0.001$). Precipitation events were short and episodic (Figure 5.2) and total precipitation inputs during the 55 day monitoring period were very low (17.4 mm). The greatest prolonged period of precipitation (DOY 224 – 227) coincided with a cold period (DOY 225 – 229) during which air temperature at all sites exhibited a marked decline (Figure 5.2). A second cold period (DOY 238 – 242) occurred at the end of the monitoring period. With the exception of these two periods, air temperature at all sites displayed no clear seasonal trends but varied on a sub-daily basis in conjunction with incoming radiation inputs.

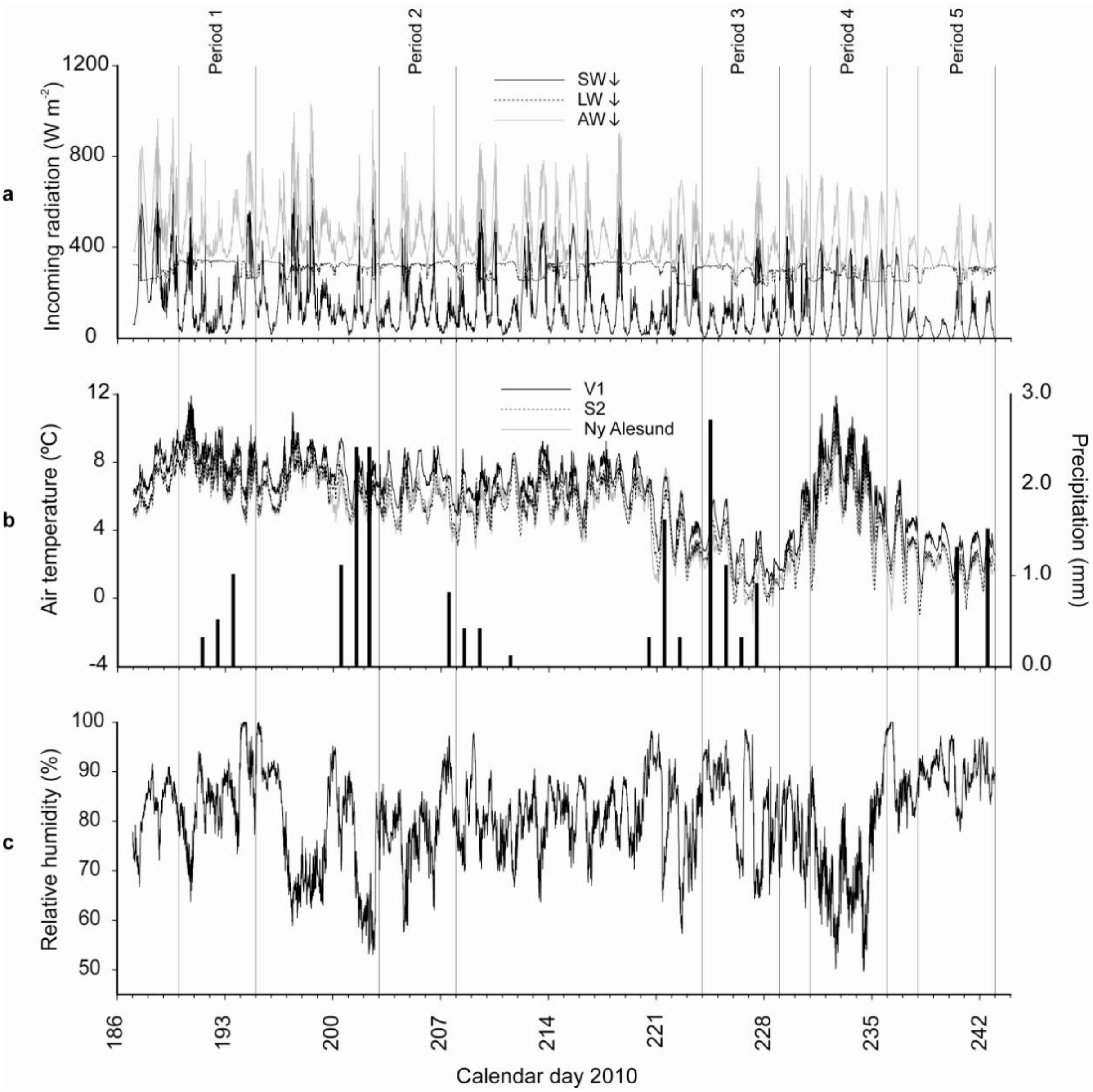


Figure 5.2: *a) Short-, long- and all-wave radiation inputs, b) air temperature and precipitation inputs, and c) relative humidity for the 2010 melt season*

Longitudinal increases in the magnitude of river discharge were observed in the Bayelva basin, although patterns of discharge were similar at all sites (Figure 5.3; Figure 5.4). Discharge was associated strongly with air temperature but not evidently affected by precipitation events. Mean discharge over the monitoring period in the Bayelva basin was 0.89 $m^3 s^{-1}$, 1.41 $m^3 s^{-1}$ and 2.69 $m^3 s^{-1}$ at V1, V2, and BR, respectively. Following peak flow at all sites on DOY 220 and 221, discharge decreased markedly and remained low for the remainder of the melt season. Mean discharge at S3 below the Stuphallet cliffs (0.04 $m^3 s^{-1}$) was markedly lower than recorded in the Bayelva basin, draining a much smaller basin area of 0.6 km^2 *cf.* 33.5 km^2 (Table 5.2). Diurnal peaks associated with high air temperature and SW↓ were superimposed on a gradual decline in discharge over the duration of the monitoring period, although the magnitude of variation (range and std. dev.) was 1-2 orders of magnitude lower for Stuphallet cliffs rivers than at sites in the Bayelva basin.

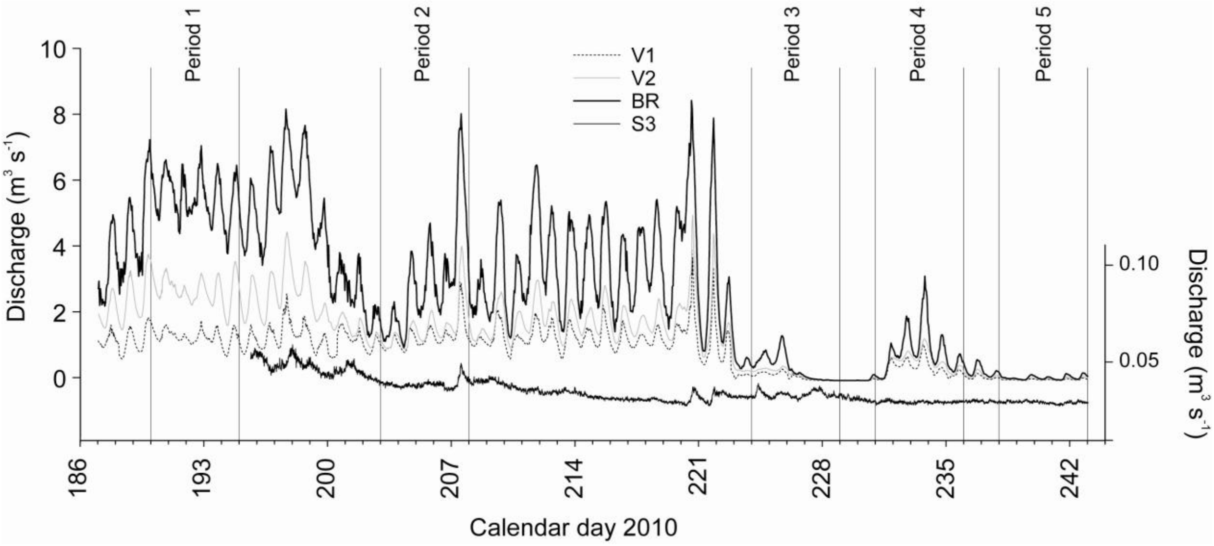


Figure 5.3: *River discharge at sites V1 (Vestre Brøggerbreen snout), V2 (Bayelva river mainstem), BR (Bayelva river mouth) and S3 (Stuphallet cliffs) for the 2010 melt season*

Table 5.2: *Descriptive statistics for air temperatures (Ny Ålesund, site V1, site S2), incoming short-, long- and all-wave radiation, precipitation, relative humidity, and river discharge (sites V1, V2, BR and S3)*

	Air temperature			*SW↓*	*LW↓*	*AW↓*	*Precipitation*	*Rel. humidity*	*Q*			
	(NÅ)	*(V1)*	*(S2)*						*V1*	*V2*	*BR*	*S3*
	°C	*°C*	*°C*	*MJ m^{-2} day^{-1}*	*MJ m^{-2} day^{-1}*	*MJ m^{-2} day^{-1}*	*mm day^{-1}*	*%*	*m^3 s^{-1}*			
*Season (days 187-242)**												
Mean (sum)	4.93	5.41	5.06	12.61	25.97	38.57	(17.00)	80.95	0.89	1.41	2.69	0.04
Max	10.90	11.17	10.80	28.38	29.21	51.79	2.70	100.00	3.94	4.99	8.44	0.06
Min	-1.50	-0.05	-0.95	2.65	22.14	28.99	0.00	49.70	0.00	0.01	0.01	0.03
Range	12.40	11.22	11.75	25.74	7.07	22.80	2.70	50.30	3.94	4.98	8.43	0.03
Std dev	2.32	2.30	2.29	6.07	1.97	5.44	0.61	9.39	0.64	1.07	2.19	0.00
Period 1 (days 190-194)												
Mean (sum)	7.36	7.77	7.40	13.86	27.88	41.74	(1.80)	84.79	1.33	2.73	5.54	-
Max	10.90	11.17	10.80	25.75	29.21	50.62	1.00	100.00	1.80	3.60	7.08	-
Min	4.30	4.83	4.47	6.57	24.87	35.78	0.00	63.80	0.73	1.66	3.83	-
Range	6.60	6.34	6.33	19.18	4.34	14.84	1.00	36.20	1.07	1.94	3.25	-
Std dev	1.18	1.14	1.14	7.60	1.78	5.94	0.42	7.38	0.22	0.41	0.67	-
Period 2 (days 203-207)												
Mean (sum)	5.62	6.32	5.67	10.66	27.51	38.17	(0.80)	79.10	1.33	1.61	3.02	0.04
Max	7.80	8.24	7.87	15.06	28.38	41.26	0.80	97.20	3.02	4.06	8.04	0.05
Min	3.30	4.26	3.88	7.51	26.21	35.54	0.00	57.60	0.85	0.95	1.00	0.03
Range	4.50	3.98	3.98	7.55	2.17	5.73	0.80	39.60	2.17	3.11	7.04	0.02
Std dev	0.98	0.95	0.92	3.05	0.81	2.42	0.36	7.31	0.40	0.65	1.60	0.00
Period 3 (days 224-228)												
Mean (sum)	1.55	2.02	1.88	8.47	24.96	33.43	(5.00)	83.46	0.21	0.29	0.51	0.03
Max	4.20	5.12	4.83	16.23	27.35	38.45	2.70	98.60	1.10	1.27	3.15	0.04
Min	-1.50	-0.05	-0.36	4.89	22.22	30.92	0.00	64.40	0.00	0.01	0.01	0.03
Range	5.70	5.17	5.19	11.34	5.13	7.53	2.70	34.20	1.10	1.26	3.14	0.01
Std dev	1.21	1.42	1.46	4.46	2.07	3.10	1.05	9.20	0.24	0.32	0.62	0.00
Period 4 (days 231-235)												
Mean (sum)	5.93	6.69	5.93	13.96	23.56	37.51	(0.00)	72.25	0.25	0.36	0.61	0.03
Max	10.90	11.17	10.80	15.62	25.26	39.09	0.00	98.70	1.10	1.27	3.15	0.03
Min	0.80	2.47	0.41	12.11	22.24	36.25	0.00	49.70	0.01	0.02	0.03	0.03
Range	10.10	8.70	10.38	3.51	3.02	2.84	0.00	49.00	1.09	1.25	3.12	0.00
Std dev	2.16	1.96	2.39	1.26	1.28	1.12	0.00	10.82	0.24	0.32	0.61	0.00
Period 5 (days 238-242)												
Mean (sum)	1.65	2.06	1.85	5.25	25.86	31.11	(2.80)	89.83	0.03	0.05	0.10	0.03
Max	3.10	3.68	3.32	7.91	27.50	32.25	1.50	97.40	0.08	0.13	0.24	0.03
Min	0.00	0.04	-0.95	2.65	24.26	28.99	0.00	78.00	0.01	0.02	0.03	0.01
Range	3.10	3.64	4.27	5.27	3.23	3.27	1.50	19.40	0.07	0.12	0.21	0.00
Std dev	0.68	0.91	0.87	2.13	1.18	1.47	0.77	3.92	0.02	0.03	0.06	0.00

** S3 discharge data from day 195 to 242*

5.3.2 Seasonal water column temperature dynamics

One-way ANOVA showed water temperature for the Stuphallet cliffs rivers was significantly warmer (mean F=6.09; p<0.05) and more variable (std. dev. F=6.47; p<0.05) than in the glacierized Bayelva basin (Table 5.3; Table 5.4; Figure 5.4; Figure 5.5). The coldest mean temperature was recorded at A1 and the warmest at S2 (Table 5.3). Diurnal cycles were evident at all sites but varied in magnitude with water temperature at glacial snouts (sites V1 and A1) exhibiting less variation than those downstream and in non-glacial systems. Meltwater sourced from Austre Brøggerbreen was colder and thermally less variable than from Vestre Brøggerbreen. The longitudinal temperature gradient observed from V1 to V2 (average rise 2.3 ºC) was punctuated by the confluence with the colder Austre Brøggerbreen river, with cooler water temperature recorded downstream at site BR (Table 5.3). Comparatively flat temperature-duration curves constructed for the entire monitoring period were indicative of thermally-constant conditions at A2, A1 and V1 while steeper curves signified greater thermal variability at other sites (Figure 5.6). These patterns were also apparent in time-series plots (Figure 5.7); water column temperatures recorded at sites in close proximity to glaciers exhibited less diurnal variability than those further downstream or in non-glacierized basins.

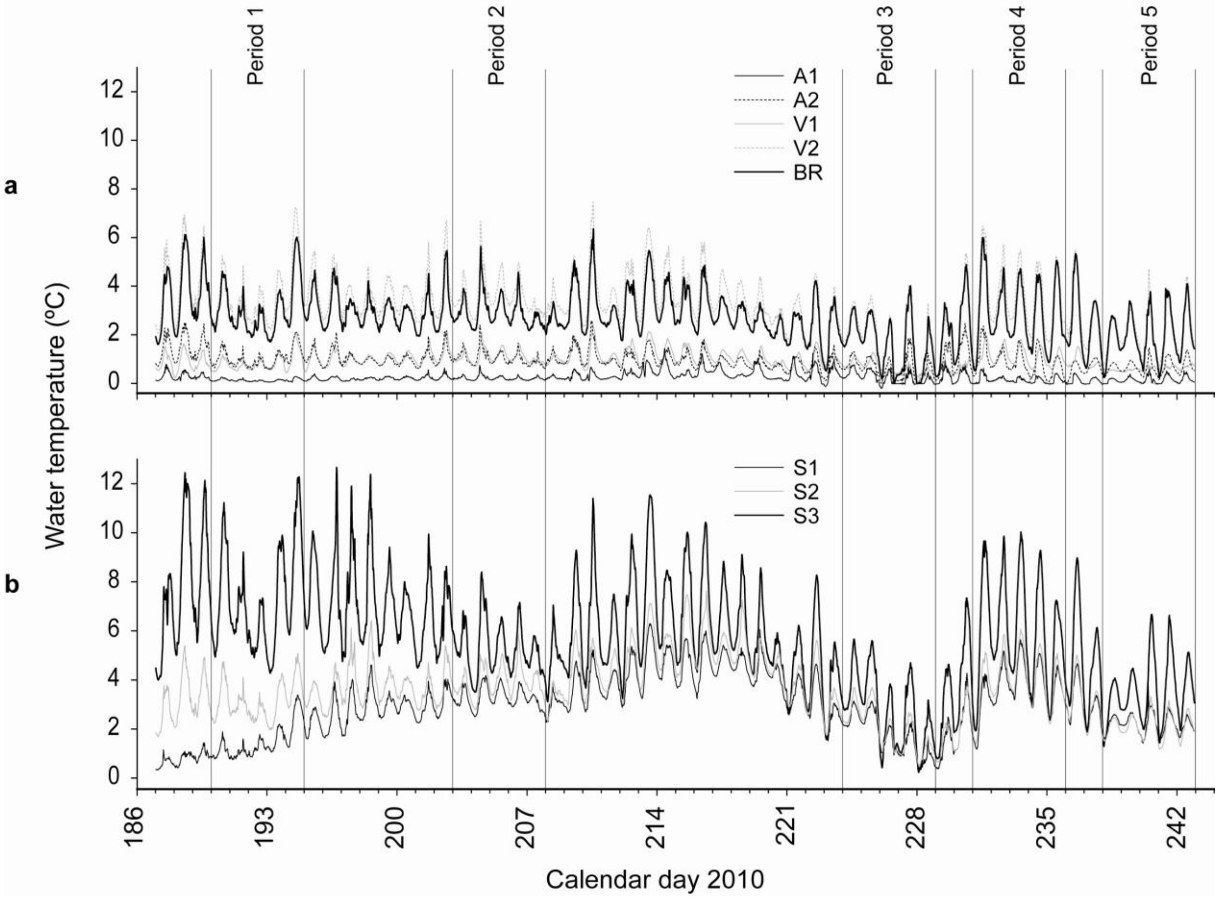


Figure 5.4: *Water column temperature at a) glacier-, and b) non-glacier-fed sites for the 2010 melt season*

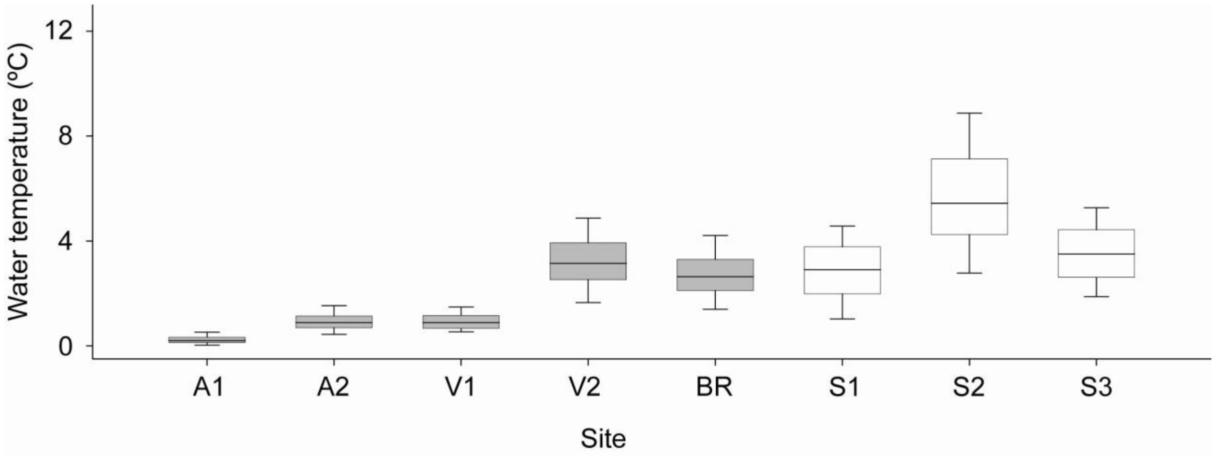


Figure 5.5: *Water column temperature boxplots for glacier-fed (shaded) and non-glacier-fed (unshaded) sites for the 2010 melt season*

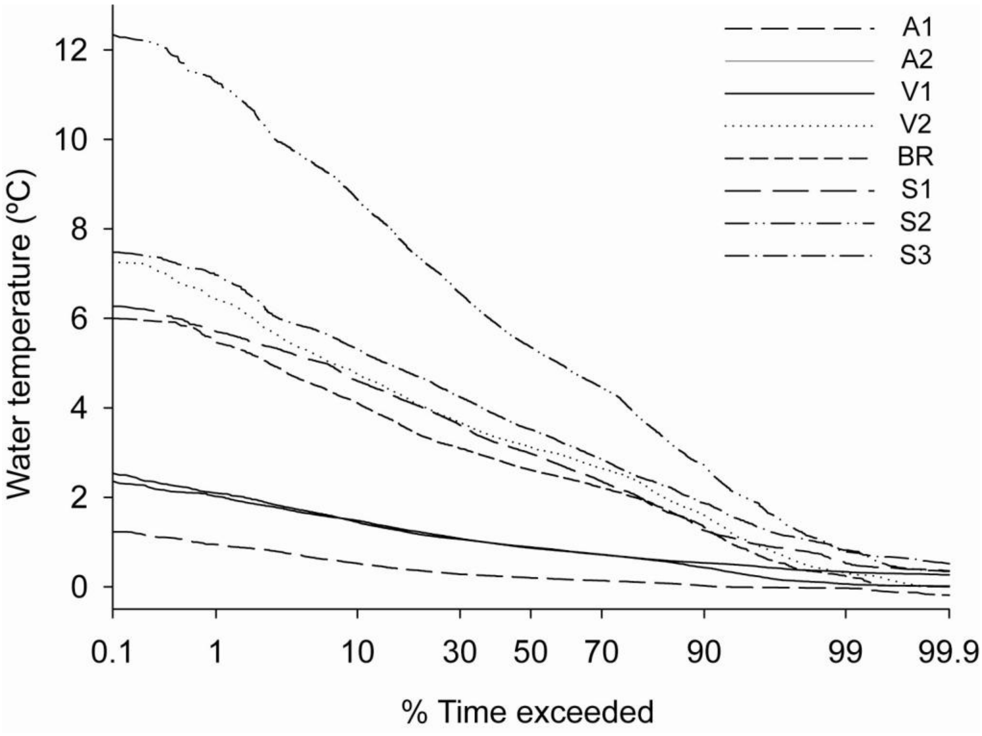


Figure 5.6: *Water column temperature-duration curves for the 2010 melt season*

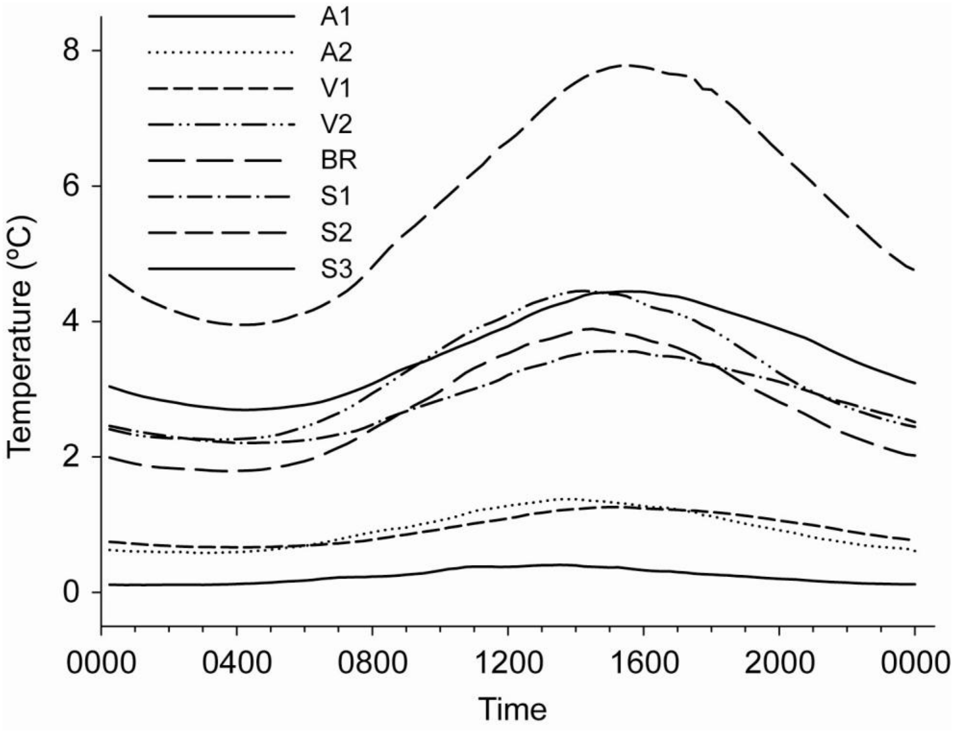


Figure 5.7: *Composite of daily water column temperature dynamics for the 2010 melt season*

Table 5.3: *Descriptive statistics for water column temperature at glacier-fed (sites V1, V2, A1, A2 and BR) and non-glacier-fed (sites S1, S2, S3) monitoring sites*

	Site							
	V1	*V2*	*A1*	*A2*	*BR*	*S1*	*S2*	*S3*
Season (days 187-242)								
Mean	0.94	3.21	0.25	0.94	2.72	2.87	5.67	3.55
Max	2.42	7.45	1.43	2.58	6.36	6.29	12.66	7.62
Min	0.23	-0.03	-0.21	0.00	-0.01	0.26	0.23	0.46
Range	2.19	7.48	1.63	2.58	6.37	6.04	12.42	7.16
Std dev	0.37	1.26	0.21	0.43	1.11	1.28	2.33	1.31
Period 1 (days 190-194)								
Mean	0.84	3.56	0.13	1.03	3.01	1.46	7.23	3.14
Max	2.02	7.25	0.29	2.11	6.01	3.39	12.28	5.08
Min	0.43	2.21	0.05	0.55	1.71	0.76	4.28	1.96
Range	1.59	5.04	0.24	1.56	4.30	2.64	8.00	3.12
Std dev	0.35	1.16	0.05	0.37	1.01	0.66	2.02	0.75
Period 2 (days 203-207)								
Mean	1.09	3.62	0.22	1.04	2.93	3.28	5.30	3.94
Max	2.09	6.67	0.74	2.39	5.63	4.17	8.40	5.91
Min	0.71	2.60	0.11	0.61	2.18	2.51	3.50	2.78
Range	1.37	4.07	0.63	1.78	3.46	1.67	4.90	3.12
Std dev	0.26	0.75	0.09	0.29	0.62	0.42	1.03	0.60
Period 3 (days 224-228)								
Mean	0.63	1.81	0.23	0.64	1.65	1.75	2.82	2.03
Max	1.27	4.08	0.77	1.88	4.01	3.15	5.66	3.69
Min	0.24	-0.03	-0.21	0.00	-0.01	0.26	0.23	0.46
Range	1.03	4.11	0.97	1.88	4.02	2.89	5.43	3.23
Std dev	0.25	1.10	0.24	0.43	1.03	0.80	1.46	0.87
Period 4 (days 231-235)								
Mean	1.09	3.49	0.15	0.91	2.86	3.67	6.03	3.97
Max	1.95	6.47	0.61	2.36	5.99	5.64	10.03	6.09
Min	0.57	0.81	-0.03	0.06	0.52	1.18	1.93	1.41
Range	1.38	5.66	0.65	2.30	5.47	4.46	8.10	4.67
Std dev	0.36	1.37	0.14	0.54	1.35	0.98	2.20	1.10
Period 5 (days 238-242)								
Mean	0.60	2.15	0.15	0.67	1.98	2.21	3.55	2.10
Max	0.84	4.71	0.59	1.46	4.10	3.90	6.67	3.33
Min	0.46	0.49	-0.04	0.15	0.23	1.43	1.29	1.17
Range	0.38	4.22	0.63	1.31	3.86	2.47	5.38	2.16
Std dev	0.09	0.93	0.14	0.29	0.98	0.40	1.27	0.48

Table 5.4: *One-way ANOVA results of water temperature descriptive statistics between glacier-fed (n=5) and non-glacier-fed (n=3) sites for the 2010 melt season*

	F	*p*
Mean	6.09	0.05
Maximum	5.09	0.07
Minimum	8.96	0.02
Range	4.38	0.08
Std dev	6.47	0.04

Local air-water column temperature correlations were consistently stronger than those based on air temperature for Ny Ålesund (Table 5.5). Correlations were weaker near glacial snouts, particularly at A1, but increased systematically downstream. The comparatively strong air-water relationships for non-glacial fed sites (S2 and S3) did not hold for S1, thus air-water column temperature correlations for the Stuphallet cliffs sites were not always higher than those in the Bayelva basin. Typically, AW↓ and SW↓ were better correlated with water column temperature than air temperature at all sites, while correlations of water column temperature with relative humidity were consistently negative ($-0.06 > r > -0.36$, $p<0.01$) (Table 5.5). Positive water column temperature-discharge correlations were observed at V1, V2 and BR in the Bayelva basin. The strongest relationship occurred at the glacial snout with a progressive decline in correlation strength downstream. A weak, albeit significant, temperature-discharge relationship was also recorded at S3. Water temperature and discharge plots indicated positive hysteresis at V1, V2 and BR where lower water temperature was associated with the falling limb of the hydrograph (Figure 5.8). No evidence of water temperature-discharge hysteresis was observed at S3.

Table 5.5: *Pearson correlation coefficients for water temperature relationships with incoming short-, long- and all-wave radiation, air temperature, relative humidity and river discharge*

Site	*r*						
	SW↓	*LW*↓	*AW*↓	*Air temperature (Ny Ålesund)*	*Air temperature (Local)*	*Relative humidity*	*Discharge*
Season (days 187-242)							
V1	0.50	-0.02 ^	0.54	0.54	0.56	-0.36	0.49
V2	0.67	0.04	0.73	0.67	0.70	-0.32	0.40
A1	0.23	0.02 ^	0.25	-0.03 *	-0.01 ^	-0.06	-
A2	0.69	-0.03 ^	0.73	0.50	0.53	-0.30	-
BR	0.70	-0.01 ^	0.74	0.60	0.63	-0.30	0.39
S1	0.19	0.02 ^	0.21	0.33	0.35	-0.28	-
S2	0.60	0.07	0.66	0.76	0.79	-0.30	-
S3	0.45	0.07	0.50	0.64	0.67	-0.33	0.09
Period 1 (days 190-194)							
V1	0.85	-0.62	0.83	0.12	0.14	0.01 ^	0.22
V2	0.88	-0.60	0.88	0.19	0.22	-0.02 ^	0.15
A1	0.79	-0.49	0.80	0.44	0.45	-0.26	-
A2	0.91	-0.64	0.90	0.23	0.28	-0.05 ^	-
BR	0.87	-0.64	0.85	0.30	0.31	-0.12 *	0.26
S1	0.77	-0.76	0.71	-0.06 ^	-0.07 ^	0.21	-
S2	0.81	-0.66	0.78	0.37	0.41	-0.19	-
S3	0.83	-0.65	0.81	0.31	0.33	-0.14	-
Period 2 (days 203-207)							
V1	0.59	-0.17	0.58	0.58	0.73	-0.63	0.11 *
V2	0.69	-0.16	0.70	0.57	0.68	-0.61	-0.19
A1	0.73	-0.19	0.73	0.25	0.38	-0.34	-
A2	0.74	-0.23	0.73	0.36	0.50	-0.51	-
BR	0.74	-0.17	0.74	0.50	0.63	-0.60	0.04 ^
S2	0.63	-0.16	0.63	0.63	0.82	-0.51	-
S1	0.51	-0.10 *	0.52	0.74	0.87	-0.49	-
S3	0.64	-0.06 ^	0.66	0.64	0.80	-0.38	0.07 ^
Period 3 (days 224-228)							
V1	0.10 *	0.47	0.28	0.67	0.71	0.27	0.82
V2	0.51	0.33	0.67	0.72	0.78	-0.12 *	0.45
A1	0.06 ^	0.56	0.27	0.75	0.76	0.25	-
A2	0.63	0.23	0.76	0.64	0.69	-0.26	-
BR	0.52	0.29	0.67	0.70	0.75	-0.16	0.44
S1	0.12 *	0.52	0.31	0.84	0.91	0.25	-
S2	0.35	0.42	0.53	0.80	0.91	0.05 ^	-
S3	0.20	0.49	0.39	0.83	0.91	0.19	0.17
Period 4 (days 231-235)							
V1	0.30	0.25	0.36	0.57	0.57	-0.38	0.69
V2	0.76	0.09 *	0.81	0.65	0.60	-0.47	0.21
A1	0.70	0.06 ^	0.75	0.48	0.43	-0.31	-
A2	0.74	0.10 *	0.80	0.65	0.60	-0.50	-
BR	0.76	0.06 ^	0.81	0.59	0.53	-0.40	0.23
S1	0.41	0.18	0.46	0.63	0.69	-0.51	-
S2	0.53	0.21	0.60	0.71	0.76	-0.53	-
S3	0.45	0.23	0.52	0.70	0.76	-0.54	0.21
Period 5 (days 238-242)							
V1	0.22	-0.19	0.17	0.10 *	0.26	-0.33	0.65
V2	0.69	-0.01 ^	0.74	0.59	0.69	-0.32	0.09 ^
A1	0.76	-0.03 ^	0.82	0.44	0.53	-0.26	-
A2	0.69	-0.01 ^	0.73	0.39	0.44	-0.32	-
BR	0.70	0.04 ^	0.77	0.64	0.74	-0.28	0.03 ^
S1	0.51	-0.07 ^	0.53	0.68	0.66	-0.44	-
S2	0.51	-0.12	0.51	0.55	0.59	-0.43	-
S3	0.52	-0.09 *	0.53	0.65	0.62	-0.41	0.15

*p<0.01 except for * p<0.05 and ^ non-significant* *- denotes not measured*
Season: n=5377 for all variables except discharge at S3 where n=4549
Subseason: n=480 for all sites

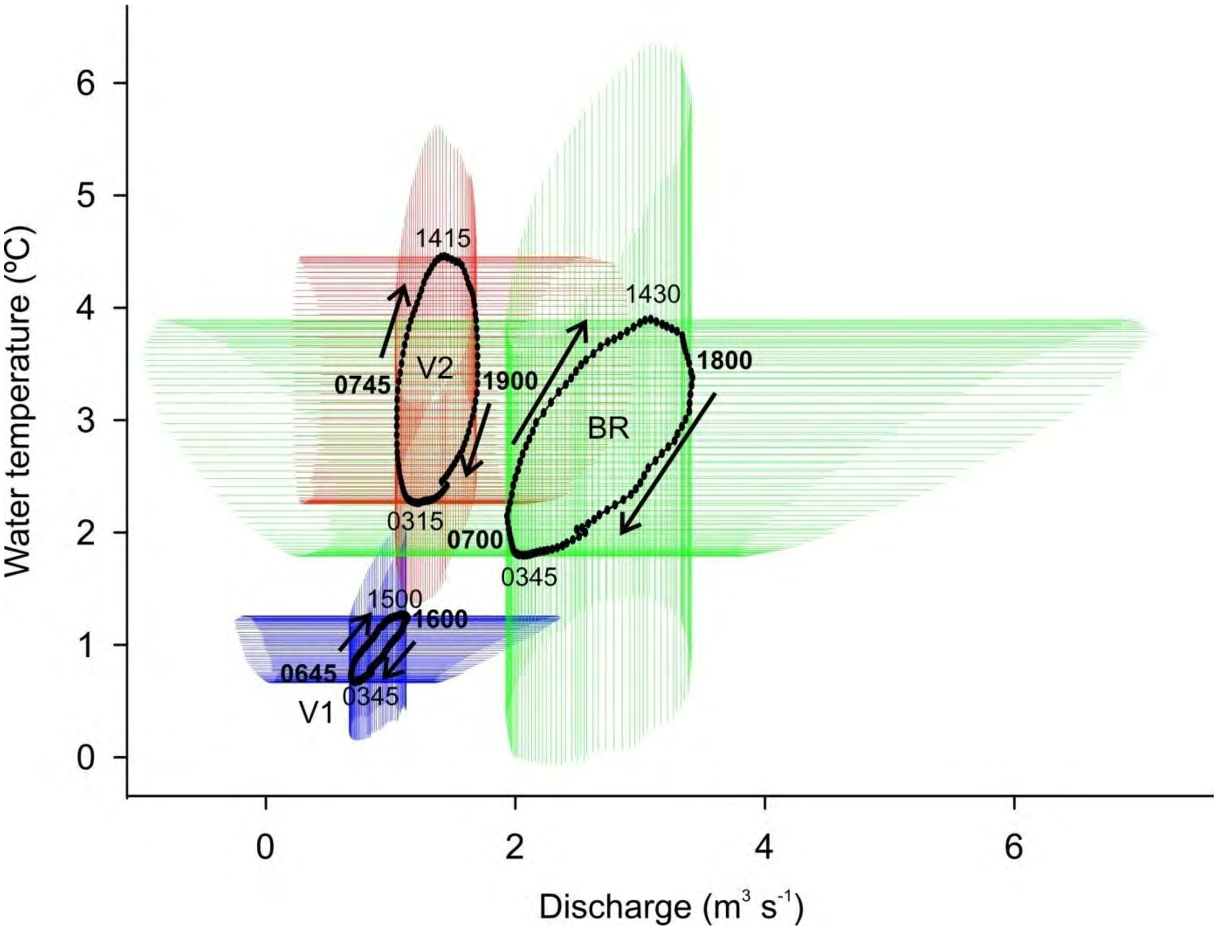


Figure 5.8: *Hysteresis patterns in mean discharge-water temperature relationships at sites V1 (Vestre Brøggerbreen snout), V2 (Bayelva river mainstem) and BR (Bayelva river mouth) for the 2010 melt season. Coloured lines indicate 1 SD around the mean. Time values represent mean maxima and minima for discharge (bold) and water temperature.*

5.3.3 Sub-seasonal context

5.3.3.1 Period 1 – Early melt season with high discharge and warm air temperature (days 190-194)

High mean daily AW↓ (41.7 MJ m^{-2} d^{-1}) and SW↓ (13.9 MJ m^{-2} d^{-1}) was observed in this period close to the summer solstice. Mean air temperature was high during this period (7.4 ºC in Ny Ålesund) with air temperature at site V1 reaching a seasonal maximum of 11.2 ºC on DOY 190. Precipitation inputs were very low (1.8 mm). Mean discharge was 1.33, 2.73 and 5.54 m^3 s^{-1} at V1, V2 and BR, respectively.

The widest range of instantaneous water column temperature between sites was found during this period, from a maximum of 12.3 ºC at S2 to a minimum of 0.1 ºC at A1. For the majority of sites, mean temperature correlated strongly with SW↓ with air-water column temperature correlations being weaker (Table 5.5). Distinct differences existed between water column temperature at non-glacier fed sites below the Stuphallet cliffs: high mean and maximum temperature at S2 contrasted with lower and less variable temperature at S1 (Figure 5.9a). Water column temperature at all monitored sites exhibited moderate positive correlations ($0.15 > r > 0.26$, $p<0.01$) with river discharge (Table 5.5).

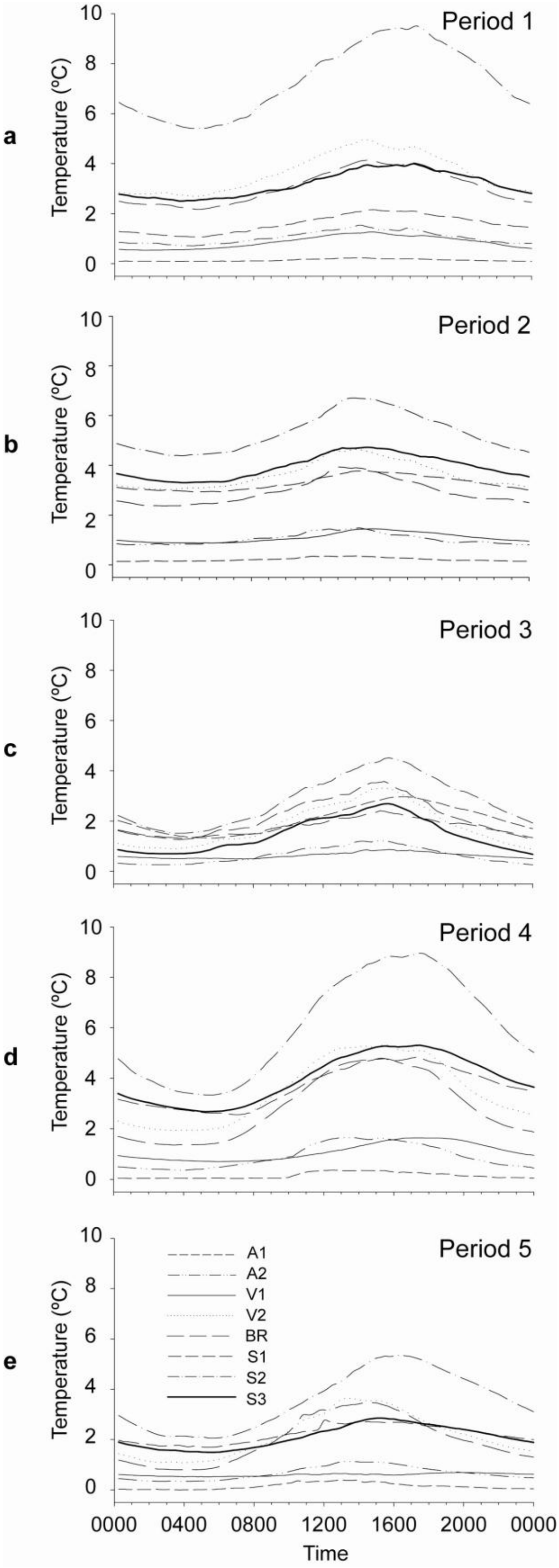


Figure 5.9: *Sub-seasonal composites of daily water column temperature dynamics for a) Period 1 (days 190-194), b) Period 2 (days 203-207), c) Period 3 (days 224-228), d) Period 4 (days 231-235), and e) Period 5 (days 238-242) for the 2010 melt season*

5.3.3.2 Period 2 – Mid-season warm period with increasing discharge (days 203-207)

Breaks in cloud cover allowed for brief peaks of high SW↓ (max 717 W m^{-2}), although average daily totals (10.7 MJ m^{-2} d^{-1}) were below the seasonal mean. Air temperature at all sites exceeded the seasonal average by approximately 0.7 ºC. A trace amount of precipitation (0.8 mm) occurred solely on DOY 207 at the end of the period. Discharge was relatively high, particularly at V1, and diurnal discharge ranges were the greatest of any selected period.

With the exception of BR and S2, mean water column temperature was higher than in Period 1 despite lower incoming short-wave radiation and air temperature during this period (Table 5.2; Table 5.3). Radiation-water column correlations were lower than in Period 1, although air-water column correlations increased (Table 5.5). A reduction in temperature at S2 and an increase at S1 reduced variability between non-glacier fed rivers compared to Period 1 (Figure 5.9b). Relatively high minimum water temperature moderated the sub-daily temperature range and contributed to less sub-daily variability at all sites (Table 5.3; Figure 5.9b).

5.3.3.3 Period 3 – Precipitation-influenced cold period with low discharge (days 224-228)

Low mean daily SW↓ (8.5 MJ m^{-2} d^{-1}) caused a significant reduction in mean AW↓ (33.4 MJ m^{-2} d^{-1}) during this period. Although precipitation over the period (5 mm) was relatively high for this arid region, a substantial fraction fell as snow in the early morning of DOY 224 and on the evening of DOY 226. Snowfall and reduced radiative inputs were associated with suppressed air temperature and reduced glacial meltwater discharge at all sites in the Bayelva basin. Declines in discharge were less marked at the non-glacier-fed S3.

The coldest water column temperatures occurred during this period (Table 5.2; Table 5.3). Daily maximum water column temperature was suppressed, thus temperature ranges were small at most sites with less inter-site variation in comparison to other periods (Table 5.3; Figure 5.9c). Despite weak or insignificant relationships between water column temperature and incoming short-wave radiation

(Table 5.5), correlations with local air temperature were significant at all sites and stronger than at any other period (0.69 > r > 0.91, p<0.01). Discharge was strongly correlated with water column temperature at sites in the Bayelva basin, particularly at V1 (r=0.82, p<0.01; Table 5.5).

5.3.3.4 Period 4 – Late season warm period with high flows (days 231-235)

Clear skies gave rise to the highest mean daily SW↓ (14.0 MJ m^{-2} d^{-1}) of any period and produced very pronounced sub-daily cycles in SW↓. Air temperature at all sites exhibited a large sub-daily range (0.4 – 11.2 ºC) and reached a seasonal maximum on day 232 (5.9 – 6.7 ºC). Compared with the previous period, discharge increased at sites in the Bayelva basin but not in the Stuphallet cliffs area.

Mean water column temperature was high during this period (Table 5.3) and large diurnal ranges occurred at most sites (Figure 5.9d). SW↓-water and air-water column temperature correlations were strong during this period (0.3 > r > 0.76, p<0.01). The water column temperature-discharge relationship at site S3 was the strongest of any sub-period (r=0.21, p<0.01; Table 5.5), although relatively weak when compared to those observed at glacier-fed sites (0.21 > r > 0.69, p<0.01).

5.3.3.5 Period 5 – Late season cold period with flow recession (days 238-242)

Mean daily SW↓ (5.6 MJ m^{-2} d^{-1}) during this period was the lowest of the season. Mean air temperature was low at all sites (1.7 – 2.1 ºC) with a relatively low diurnal range (< 4.3 ºC) in comparison with other periods. Discharge was the lowest of any sub-period at all sites and showed limited sub-daily variation. Two precipitation events (days 240 and 242) resulted in a total of 2.8 mm of rainfall during the period (Table 5.2).

Mean water temperature was among the coldest recorded during the monitoring period (Table 5.3). Temperature variability (range and std. dev.) was low at most sites, although diurnal cycles remained visible but less pronounced than during Periods 1 and 4 (Table 5.3; Figure 5.9e). SW↓-water and air-water column temperature correlations were generally strong (*r*>0.50, *p*<0.01 for most sites) with the exception of V1 where water column temperature, in contrast to earlier periods, appeared to be decoupled from meteorological variables. Water column temperature-discharge relationships were positively correlated at V1 and S3 (*r*=0.15 and 0.65, respectively, *p*<0.01) but insignificant at V2 and BR (Table 5.5).

5.4 Discussion

The high-resolution data generated by this research have provided new insight into spatiotemporal water temperature patterns within Arctic river basins, and demonstrated the strong influence of hydroclimatological conditions and water source on river temperature regimes. Clear differences in river thermal dynamics are evident across this region of Svalbard: significantly warmer and more variable water temperature regimes occurred in non-glacierized basins where flow was maintained principally by meltwater derived from seasonal snowpacks and shallow groundwater, while mean water temperature at glacier-fed sites was lower and more thermally-stable, particularly near the glacier terminus.

5.4.1 Controls on seasonal water temperature dynamics

Mean water temperature, and the relative magnitude of daily thermal variation, were broadly similar to Arctic systems at lower latitudes, although the maximum summer water temperature was approximately 9 ºC lower in Svalbard (79°N) than in Alaskan rivers located between 66 - 69 ºN (Irons and Oswood, 1992; Chikita *et al.*, 2010). Water temperature was lower in all of the study rivers when compared with those recorded in alpine environments (Uehlinger *et al.*, 2003; Brown *et al.*, 2005; Cadbury *et al.*, 2008). This finding reflects the reduced magnitude of atmospheric energy inputs (characterised herein by AW↓, SW↓ and air temperature) and strong seasonality experienced in high-latitude regions. Changes in SW↓ accounted for much of the variability in AW↓ and appeared to be the primary driver of water temperature at all sites (*cf.* Joly and Brossard, 2007), and also of glacier meltwater generation in the Bayelva basin as inferred from discharge associations (*cf.* Hodgkins, 2001). The role of SW↓ as a major river heat source is well-documented (Hannah *et al.*, 2008). This is also consistent with the predominantly negative LW↓- and relative humidity-water temperature relationships, because both variables can be viewed a surrogates for cloud cover and SW↓ typically decreases under overcast conditions (Oke, 1987).

In general, water temperature increased longitudinally downstream of the glacier terminus with a 0.7 ºC km^{-1} and 2.3 ºC km^{-1} temperature increase from Austre Brøggerbreen and Vestre Brøggerbreen, respectively. Such longitudinal rises in water temperature are typical of glacial river systems; previously documented increases have ranged from 0.4 – 0.6 ºC km^{-1} (Uehlinger *et al.*, 2003; Cadbury *et al.*, 2008), although on rare occasions have been recorded as high as 7 ºC km^{-1} (Brown *et al.*, 2005; Brown and Hannah; 2008). Buried ice in the proglacial area (Engeset and Weydahl, 1998) could suppress rises in water temperature by modifying riverbed thermal gradient and associated heat exchange (Hannah *et al.*, 2008). High suspended sediment loads in these glacier-fed rivers (Hodson *et al.*, 1998a) may also affect water column temperature by modifying albedo and increasing short-wave absorption (Han*,* 1997*;* Richards and Moore, 2011). Additionally, differences in the rate of temperature increase between the two Svalbard glacier-fed rivers can be explained in part by basin-specific features. Firstly, river discharge between A1 and A2 was confined to a narrow, incised channel that limited exposure to atmospheric influences. In contrast, meltwater from Vestre Brøggerbreen spread laterally across the floodplain between V1 and V2, hence width:depth ratios were high and potential for warming increased (Webb and Nobilis, 1994; Brown and Hannah, 2008). Furthermore, discharge at V1 comprised only a portion of bulk meltwater originating from Vestre Brøggerbreen. Additional lateral drainage channels converge in a shallow lake below the glacial snout before forming a confluence with waters from V1, resulting in increased discharge at V2. Pro-glacial lakes have been identified previously as modifiers of longitudinal temperature profiles in glacial rivers (Burgherr and Ward, 2001; Milner *et al.*, 2001; Moore *et al.*, 2009) with outflow temperatures typically greater than those found in river systems without lakes (Hieber *et al.*, 2002).

Mean water temperature at site BR (2.7 ºC) was 0.5 ºC cooler than at V2, despite being situated 1500 m further downstream. Tributary inputs can interrupt temperature gradients in river systems (Milner *et al.*, 2001; Knispel and Castella, 2003; Brown and Hannah, 2008). In the Bayelva basin, a reduction in mean water temperature between V2 and BR was likely to be attributable to the confluence of the colder discharge from Austre Brøggerbreen with warmer waters from Vestre Brøggerbreen below V2. Additional inputs of extra-glacial waters to the Bayelva river, and their capacity to modify thermal regimes, were considered to be negligible following peak snowmelt in early July given the shallow groundwater system (Roth and Boike, 2001), low basin storage and minimal precipitation that occurred during the monitoring period.

Positive correlations between water column temperature and river discharge were strong with clear diurnal cycles observed in both variables for a greater part of the monitoring period. The temperature-discharge relationship was particularly strong near the snout of Vestre Brøggerbreen where a small and narrow hysteresis loop reflected relatively low variability in both variables. This is a characteristic feature of glacier-fed rivers (Uehlinger *et al.*, 2003) because discharge and water temperature dynamics in these systems are influenced strongly by radiation receipt and air temperature (Hock, 2005; Hannah *et al.*, 2007; Chikita *et al.*, 2010), thus potential diurnal increases in water temperature can be offset by corresponding rises in discharge. The larger, more vertically elongated hysteresis loop at V2 reflected higher variability in water temperature over a relatively low discharge range, and *vice versa* at BR. Stronger temperature-flow hysteresis effects such as these have not been documented previously for glacier-fed rivers, although similar air-water temperature hysteresis relationships are known to occur in larger temperate rivers influenced by inputs from snowmelt which can lead to reduced water temperature despite high air temperature (Webb and Nobilis, 1994; van Vliet *et al.*, 2011). Diurnal differences in timing between peak water temperature (early afternoon) and discharge (early evening) suggest that water temperature responded rapidly to variation in atmospheric controls. However, discharge lagged temperature due to delayed drainage from the Vestre Brøggerbreen glacier (Hodson *et al.*, 1998b). Weaker water column temperature-discharge relationships further downstream in the Bayelva basin indicate a lack of additional meltwater inputs downstream of glacial snouts, and a progressive differentiation of discharge and water column temperature dynamics in response to sustained atmospheric exposure.

Non-glacier fed rivers below the Stuphallet cliffs were warmer and exhibited greater thermal variation (diurnal and seasonal) than those in the Bayelva basin. Similar positive associations between mean water temperature and the daily temperature range have also been recorded for Arctic Alaskan rivers (Irons and Oswood, 1992). High mean water temperature and strong diurnal thermal cycles in non-glacier fed rivers are indicative of water derived from shallow hillslope groundwater systems (Kobayashi *et al.*, 1999; Brown *et al.*, 2005). Unlike some rivers in Alaska where flow is often maintained on a perennial basis by spring discharge (Parker and Huryn, 2011), sub-permafrost aquifer discharge on the Brøgger peninsula is thought to be restricted by permafrost aquitards and limited largely to the temperate parts of glacier base areas (Haldorsen *et al.*, 2002) and the exit of a now abandoned coal mine near Ny Ålesund (Haldorsen *et al.*, 2010). Therefore, deeper groundwater-surface water interactions are likely to be limited in the Stuphallet cliffs area, resulting in saturated areas of supra-permafrost standing water where surface flow velocities are retarded further by abundant vegetation growth. These waters have strong potential for greater atmospheric equilibration. The majority of river discharge in this area is sourced from snowpack meltwater that reaches rivers by near-surface pathways in the shallow alluvial zone rather than as overland flow (supported by field observations). The seasonal decline in discharge at S3 supports this premise as it indicates a decline in source water after snowpack depletion (Malard *et al.*, 1999; Brown *et al.*, 2006a). Additionally, relatively high mean electrical conductivity values recorded at S2 are indicative of solute enrichment acquired during flow through shallow alluvial groundwater systems (Malard *et al.*, 1999; Hodson *et al.*, 2002). Greater residence time in the alluvial zone accounts for the weak water column temperature-discharge relationship observed at S3, and also for the higher water column temperature at S2 and S3 due to the prolonged exposure of soil water to atmospheric influence. Cooler mean water temperature at S1 than at other non-glacial sites may be explained by differences in residence times: minimal soil development at Kvadehuksletta towards the western end of the Brøgger peninsula (Hallet and Prestrud, 1986) facilitates rapid runoff of meltwater from cold snowpacks, resulting in a thermal regime bearing closer resemblance to the glacier-fed Bayelva sites compared with areas where organic ‗soil' layers are deeper and atmospheric exposure time is greater.

5.4.2 Controls on sub-seasonal water temperature dynamics

Sub-seasonal patterns in river thermal dynamics appeared to be related largely to changes in hydroclimatological variables and relative water source contributions to bulk flow. The weaker correlation of water temperature with both air temperature and river discharge early in the melt season was likely related to the input of cold meltwater from valley snowpacks (Webb and Nobilis, 1994; Brown *et al.*, 2006a). River temperature regimes became more responsive to climatological forcing (particularly SW↓) following the loss of river network snowpack cover by the end of July, coinciding with the findings of Malard *et al.* (1999). Cooler water temperature at all sites in late August coincided with pronounced decreases in SW↓ following the first sunset of summer and the end of the polar day.

Discharge patterns at BR from mid-July to late-August tracked those at V1 and V2 to a greater extent than earlier in the melt season, suggesting a relative (but not absolute) increase in the contribution of glacial meltwater to flow in the lower Bayelva river in late summer. While precipitation inputs were minimal throughout the monitoring season, sub-periods with higher precipitation coincided with lower water and air temperatures. Discharge trends were not affected by precipitation events as documented previously in other glacial basins and elsewhere (Brown *et al.*, 2005; Brown & Hannah, 2007; Chikita *et al.*, 2010), most likely because the volume of precipitation during each rainfall event in Svalbard was much lower than typically experienced in alpine regions (Brown and Hannah, 2007). Such low precipitation inputs are unlikely to have directly induced cooler river temperature by heat advection as noted by Brown and Hannah (2007). However, the presence of cloud cover associated with precipitation events reduces atmospheric energy receipt, thus influencing river thermal regimes (Mohseni and Stefan, 1999).

5.5 Conclusion and implications

Research in this chapter investigated water temperature dynamics, and explored links to controlling hydroclimatological factors, in High Arctic river basins located in north-west Svalbard. High spatiotemporal heterogeneity of water column temperatures were related to river water source and prevailing meteorological conditions (Hypothesis I), and also basin-specific features (specifically channel geomorphology and the presence of lakes), river discharge, and (in the case of glacier-fed rivers) distance from source. A conceptual model summarizing the relative influence of these factors on river temperature over a longitudinal gradient is presented in Figure 5.10a. Non-glacier fed rivers were more hydrologically stable and generally exhibited warmer temperature and higher thermal variability than those fed by glacial meltwater (supporting Hypothesis II). Mean water temperature and the relative magnitude of daily thermal variation in all rivers exhibited similar patterns to those in Alaska (Chikita *et al.*, 2010; Irons and Oswood, 1992), but were markedly lower than those recorded in alpine environments (e.g. Brown *et al.*, 2005; Cadbury *et al.*, 2008) due to inferred runoff interaction with permafrost and the relatively low magnitude of incoming solar radiation and cooler air temperatures associated with the high-latitude geography of Svalbard (Hypothesis III). The key differences between Arctic and alpine river thermal regimes at seasonal and subdaily temporal scales are outlined in Figure 5.10b

Projected warming in the Arctic during the 21st Century (Holland and Bitz, 2003; Anisimov *et al.*, 2007; Serreze *et al.*, 2009) is expected to produce an initial increase in glacial meltwater generation in large systems (Fleming and Clarke, 2003; Milner *et al.*, 2009), followed by a long-term decline in discharge as glaciers waste and recede (Barnett *et al.*, 2005). Furthermore, model projections suggest a future reduction in seasonal snow cover and greater number of snow-free days each year (Anisimov *et al.*, 2007), and increased diffuse groundwater discharge associated with an increase in the depth of seasonal permafrost thawing (Walvoord and Striegl, 2007). Ultimately, such changes will reduce the absolute and relative contributions of ice and snow meltwaters to total river flow. The results of this chapter suggest that the coldest and most thermally-stable conditions are found in close proximity to glacial snouts. Thus, future changes in Arctic river temperature regimes are expected to comprise of increased mean water temperature and greater thermal heterogeneity as the proportion of river discharge sourced from meltwater declines (Hannah *et al.*, 2007; Brown *et al.*, 2009; Milner *et al.*, 2009). The greatest increases are likely in late summer and early autumn following melting of seasonal snowpacks (Webb and Nobilis, 1994).

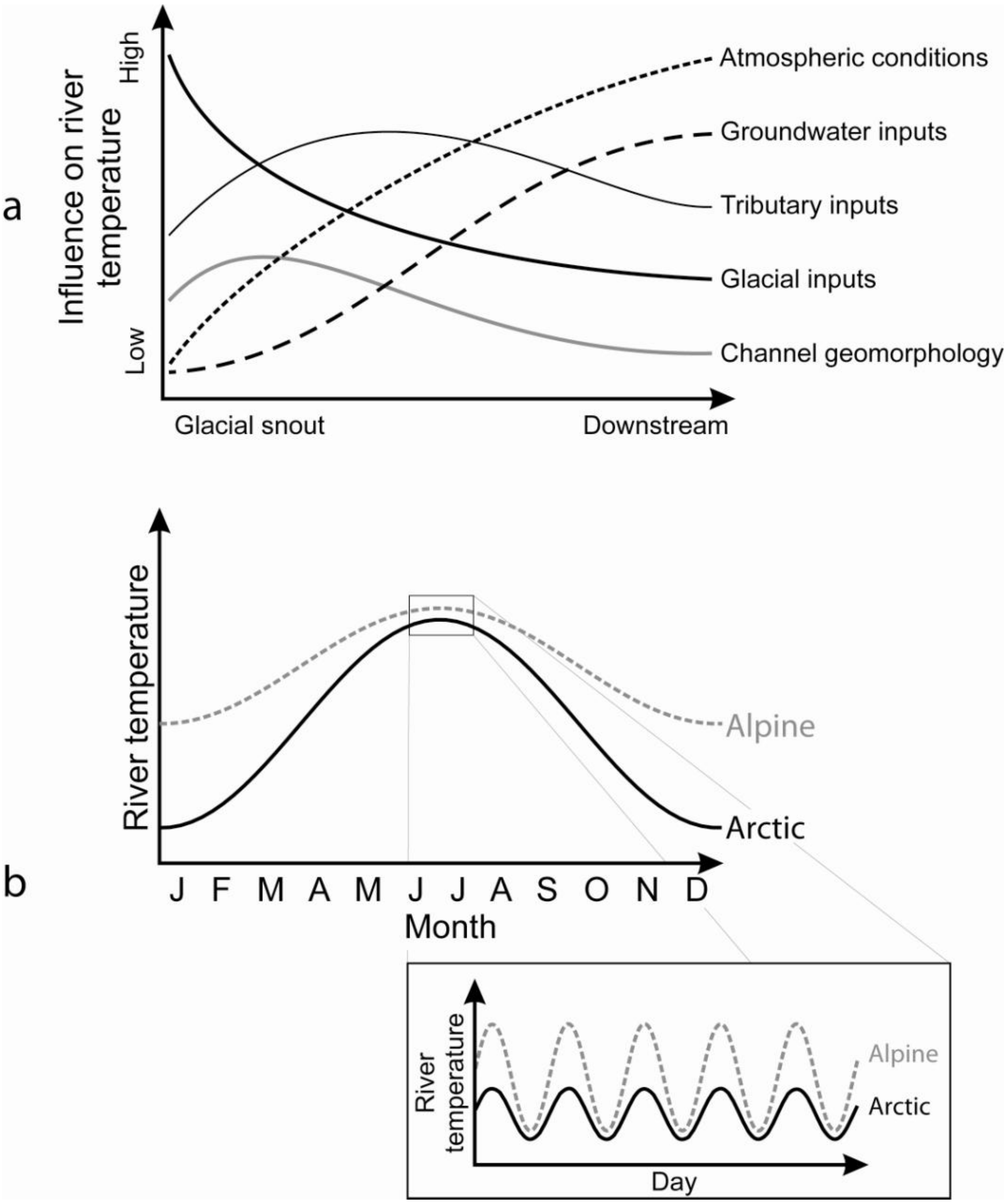


Figure 5.10: *Conceptual models of a) the relative influence of drivers of water temperature over a longitudinal gradient in an Arctic river basin, and b) key differences between Arctic and alpine river thermal regimes at seasonal and sub-daily temporal scales*

5.6 Chapter summary

Research in this chapter used a high-resolution temperature monitoring network to characterise temporal and seasonal water temperature dynamics in Svalbard rivers and assess the primary drivers of thermal variability in these systems. Future changes in high-latitude river temperature regimes associated with forecasted climate change were also discussed. River thermal shifts could have implications for rates of ecosystem functioning (e.g. nutrient uptake, leaf litter decomposition) in high-latitude aquatic ecosystems (Friberg *et al.*, 2009; Rasmussen *et al.*, 2011) and affect species diversity and abundance in benthic communities (Milner *et al.*, 2001, Brown *et al.*, 2007a; Jacobson *et al.*, 2012). These are considered in Chapters 6 and 7, respectively.

CHAPTER 6

IMPACT OF CHANGING HYDROLOGY ON NUTRIENT UPTAKE IN ARCTIC RIVERS

6.1 Introduction

Reductions in glacier mass and seasonal snow cover in response to atmospheric warming (Hinzman *et al.,* 2005; Anisimov *et al.*, 2007) are expected to affect Arctic hydrology in the near future as the proportion of meltwater contributing to total river discharge decreases (Smith *et al.*, 2007; Walvoord and Streigl, 2007). In addition, concerns have been raised that melt-induced changes in the structure and extent of near-surface permafrost systems will mobilise large pools of nitrogen and carbon stored in tundra ecosystems which could be flushed into rivers (Petrone *et al.*, 2006; Holmes *et al.*, 2008; Frey and McClelland, 2009). However, climate warming and a shift towards groundwater-dominated systems are likely to be associated with warmer water temperature, less variable hydrological regimes and higher biological activity which could enhance aquatic nutrient uptake due to increased biotic assimilation in these systems (Prowse *et al.*, 2006; Rasmussen *et al.*, 2011; Blaen *et al.*, *in press*).

A review of the literature (Chapter 2) showed that there is a paucity of information regarding controls on nutrient cycling in high-latitude rivers and thus the evidence base for projection of future changes in nutrient uptake remains uncertain. There is now a need for process-based research to quantify rates, and driving mechanisms, of nutrient uptake in Arctic river systems. The aim of this chapter is to characterise water sources, hydrological and biogeochemical characteristics, and rates of N and C uptake across a gradient of meltwater to groundwater influence as an analogue to understand how changing hydrology and environmental conditions may alter nutrient cycling in high-latitude regions. Based on the contemporary understanding that hydrological and biochemical conditions influence nutrient processing dynamics in headwater rivers, the following hypotheses were tested:

I. decreased meltwater flow contribution associated with warmer water temperature and more stable channel morphologies will support higher rates of autotrophy and microbial activity (Parker and Huryn, 2011; Blaen *et al.*, *in press*);

II. decreased meltwater flow contributions associated with altered habitat conditions and increased transient storage will increase nutrient uptake velocities and rates of retention (Zarnetske *et al.*, 2007; Rasmussen *et al.*, 2011); and

III. the addition of labile DOC stimulates N uptake by releasing N assimilation from C limitation (Johnson *et al.*, 2009).

6.2 Methodology

6.2.1 Study sites

The study was undertaken in the Kongsfjorden area of north-west Spitsbergen. Twelve first-order rivers were chosen to represent a gradient of water source inputs from meltwater and shallow active layer hillslope groundwater reservoirs (Figure 6.1). Sites X1-X6 were studied in 2011 and sites Y1-Y6 were studied in 2012. Study reaches in each river were 60-70 m long, confined to a single channel with no tributary inputs, and selected to minimise longitudinal variation in gradient and discharge.

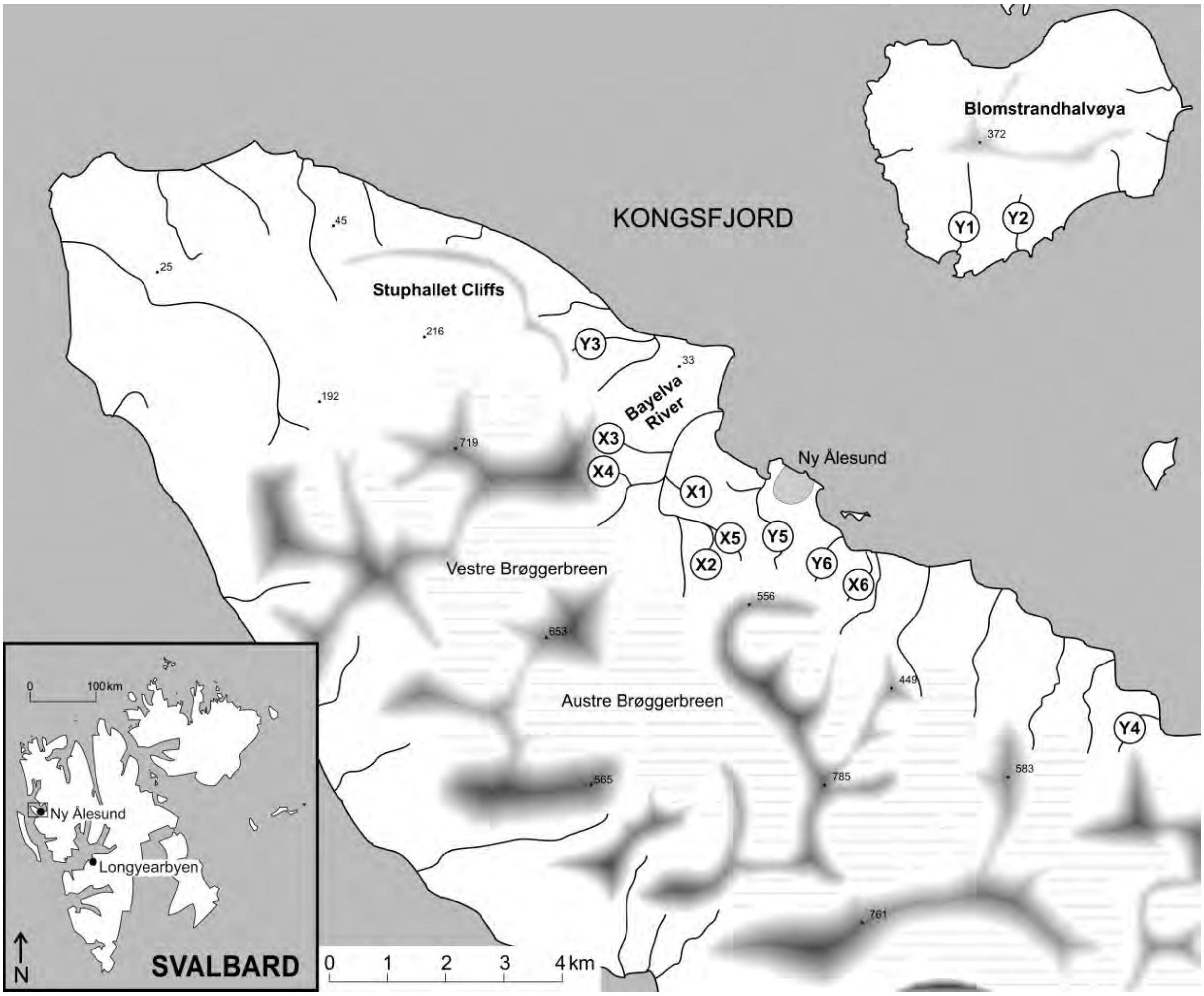


Figure 6.1: *Map of study area showing sampling sites, approximate river courses, relief (shaded), spot height measurements in meters asl, and glacier cover (dashed areas).*

6.2.2 Sampling framework

Water source contributions to river flow, transient storage metrics and hydrological and biogeochemical characteristics were quantified at each study site. These data were combined with data from short-term nutrient uptake experiments to understand the major physical and biological controls on nutrient uptake in these Arctic river systems. Here, _uptake‘ is defined as the net removal of nutrients from the water column, although the methodology employed does not distinguish the specific mechanisms responsible for this transformation.

6.2.3 Water source quantification

A two-component end-member mixing model was used to separate river discharge at each study site into two conceptual source components: (i) meltwater derived from snow and glacial ice ablation, and (ii) shallow hillslope groundwater transmitted to the river channel via subsurface flow through the active layer, following the methodology outlined in Chapter 4.

6.2.4 Nutrient uptake

Nutrient uptake was measured once at each study site during July / early August 2011 or July 2012. River levels were stable and minimal precipitation (< 2mm) occurred in the two days preceding each injection. Water samples were collected at 10 m intervals along each study reach prior to experimentation to characterise ambient solute concentrations. Short-term additions of nitrogen (as NH_4^+ and NO_3^-), phosphate (PO_4^{3-}) and acetate, plus a conservative tracer (NaCl) were used to measure nutrient uptake rates following field methods outlined in Tank *et al.* (2006). Estimates of ambient concentrations required to calculate enrichment factors were based on field data acquired during a pilot study in 2010, while discharge was estimated from velocity-area measurements obtained approximately 24 hours before each injection. Target enrichment levels above ambient concentrations were 15 µg l^{-1} NH_4^+, 20 µg l^{-1} NO_3^-, 30 µg l^{-1} PO_4^{3-} and 100 µg l^{-1} acetate. The target increase in electrical conductivity (EC) was 5-20 µS cm^{-1}. It is recognised that raising nutrient levels above ambient conditions can result in overestimation of uptake lengths (Mulholland *et al.*, 2002). However, the modest enrichment factors employed in this study suggest that measured uptake lengths are unlikely to deviate substantially from ambient lengths (Payn *et al.*, 2005; Rasmussen *et al.*, 2011).

Five nutrient additions were performed at each site: NO_3^-, NH_4^+, PO_4^{3-} and acetate were initially added separately prior to a further addition of NH_4^+ and acetate in unison. A solution containing either NH_4Cl, $NaNO_3$, Na_2HPO_4, CH_3COONa, or NH_4Cl and CH_3COONa, plus NaCl as a conservative tracer measured as Cl^-, was pumped continuously into a well-mixed area at the upper end of each reach using a Watson-Marlow peristaltic pump with a flow rate of 0.12 l min^{-1}. EC was monitored every minute (2011) or every 20 s (2012) at the upstream and downstream end of each study reach. EC values were used to calculate discharge during nutrient release and assess the time for nutrient concentrations to stabilise. Water samples were collected at 10 m intervals along each study reach once EC had reached asymptote (around 40 minutes after starting each injection).

Uptake parameters were derived from standard protocols as described by Tank *et al.* (2006) based on the longitudinal decrease in nutrient concentration through each study reach. Nutrient concentrations were divided by Cl^- concentration to correct for minor changes in discharge throughout the study reach. Nutrient uptake length (S_w, m) was calculated as the inverse of the slope of the line (k) relating background- and discharge-corrected nutrient concentration to distance downstream of the injection point:

$$\ln N_x = \ln N_0 - kx$$

where N_0 is nutrient concentration at the addition site, N_x is nutrient concentration at x meters downstream of the addition site, and k is the per-meter uptake rate (Bernhardt and Likens, 2002; Tank *et al.*, 2006). Linear decreases in nutrient concentration through each reach were considered significant where $p<0.05$.

Uptake lengths can be influenced strongly by stream discharge (Hall and Tank, 2003); therefore uptake velocity (V_f, m min^{-1}) was calculated to allow for discharge-independent comparisons of uptake from the water column:

$$V_f = Qk/w$$

where Q is discharge (m^3 min^{-1}) and w is the mean wetted channel width. Wetted channel widths were measured every 3 m along each study reach at the end of each field day to avoid disturbing bed sediments. The areal uptake rate (U, mg m^{-2} min^{-1}) for each nutrient was calculated as:

$$U = V_f \mathrm{N_a}$$

where N_a is the mean ambient nutrient concentration prior to release.

6.2.5 Solute transport modelling

Transient storage represents the temporary retention of solutes in eddies, side pools, vegetation or hyporheic zones that are stationary relative to faster moving water in the main channel (Runkel, 1998). Transient storage parameters were estimated for rivers from EC data using a one-dimensional advection-dispersion model (OTIS-P) to provide an integrated measure of hydrological retention in each reach. The model solves the following coupled differential equations (Bencala and Walters, 1983):

$$\frac{\partial C}{\partial t} = -\frac{Q}{A}\frac{\partial C}{\partial x} + \frac{1}{A}\left(AD\frac{\partial C}{\partial x}\right) + \alpha(C_s - C)$$

$$\frac{\partial C_s}{\partial t} = \alpha\frac{A}{A_s}(C - C_s)$$

where A is the main channel cross-sectional area (m²), A_s is the storage zone cross-sectional area (m²), C is the main channel solute concentration (µS cm^{-1}), C_s is the storage zone solute concentration (µS cm^{-1}), D is the dispersion coefficient (m² s^{-1}), Q is river discharge (m³ s^{-1}), t is time (s), x is distance (m) and α is the storage zone exchange coefficient (s^{-1}). Given a suite of initial boundary conditions for the head of each reach, the model adjusts parameters iteratively to produce a least-squares best fit to the downstream conservative tracer breakthrough curve measured as EC in the field (Figure 6.2). The Damkohler I number (*DaI*) was calculated to evaluate the reliability of the OTIS-P model outputs (Edwardson *et al.*, 2003):

$$DaI = \frac{\alpha\left(1 + \frac{A}{A_s}\right)L}{v}$$

where L is the reach length and v is the average velocity in that reach. Uncertainty in estimates of parameters are lowest when DaI values approach 1 and are reliable to values <10 (Wagner and Harvey, 1997). Mean hydraulic residence time in the storage zone ($T_{stor} = A_s/A\alpha$) was calculated from model parameters. The ratio of the storage zone cross-sectional area to the main channel cross-sectional area (A_s/A) was calculated to allow for comparison of relative transient storage zone size between rivers (Morrice *et al.*, 1997).

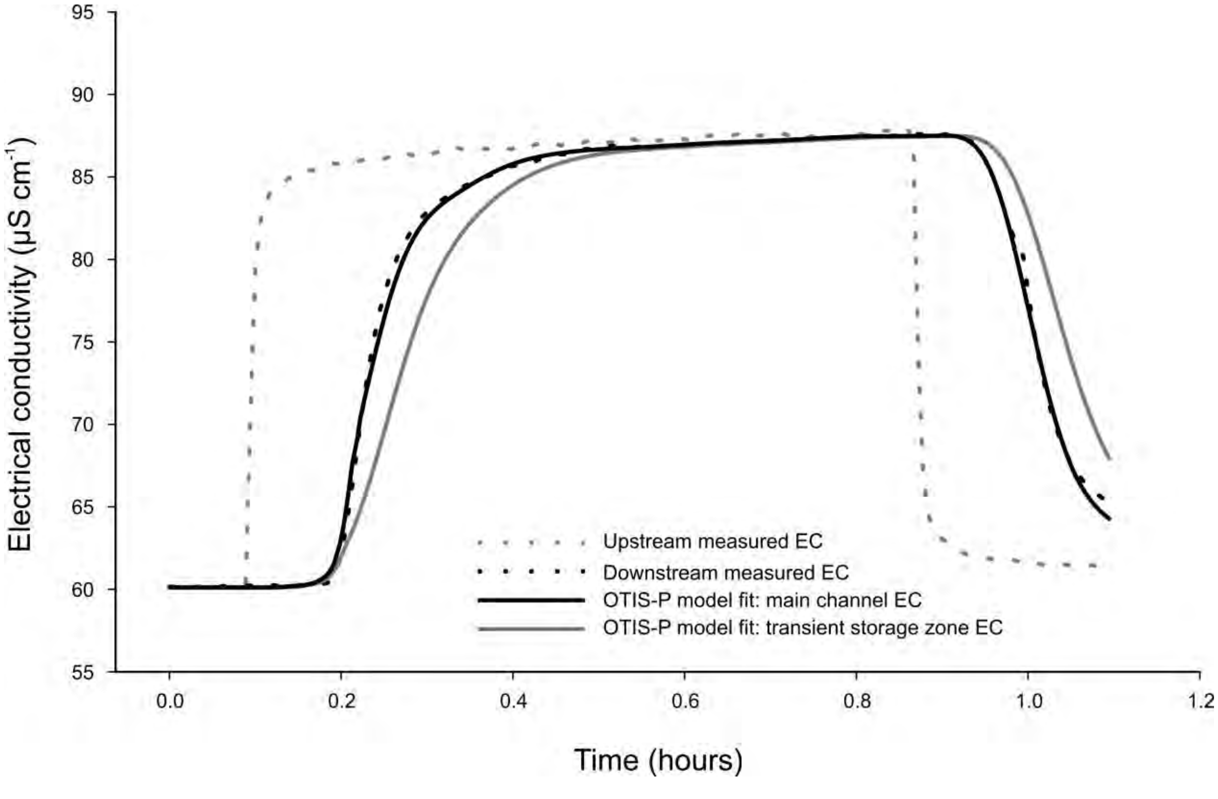


Figure 6.2: *Solute concentration curves as measured by EC for site Y4 and OTIS-P model outputs for the main channel and transient storage zone.*

6.2.6 Environmental variables

Several ancillary variables for each study reach were recorded in conjunction with nutrient uptake measurements to examine potential environmental controls driving nutrient demand. Mean water temperature during each injection was calculated from 15 min measurements at the lower end of each reach. Incoming short-wave radiation was measured in Ny Ålesund. Suspended sediment concentration (SSC), channel stability (Pfankuch Index), chlorophyll *a* concentration and sediment respiration were measured in each reach following nutrient injections. Dissolved oxygen was not measured because field testing indicated supersaturation of DO is common to small rivers in the Kongsfjord region of Svalbard.

6.2.7 Data analysis

Differences in nutrient concentration between snow and groundwater samples for each study river basin were assessed using independent samples *t*-tests. Potential associations between water source contributions, physicochemical and biological environmental variables, and uptake parameters (V_f and U) were assessed using Spearman‘s rank correlation analysis. Non-parametric correlation analysis was used due to the relatively small size of the dataset and the potential for non-linear relationships (Zar, 2010). The significance of differences in V_f between nutrient types were tested for by one-way ANOVA followed by a Tukey‘s *post-hoc* comparisons test. To assess the influence of DOC on NH_4^+ uptake and *vice versa*, demand ratios were calculated as the ratio of V_f during the simultaneous release of NH_4^+ and acetate to the V_f of each nutrient during individual releases. A ratio >1 indicates that V_f during the simultaneous release was greater than during the individual release, a ratio=1 indicates no difference in V_f, and a ratio <1 indicates a decrease in V_f for the simultaneous release relative to the individual release (Johnson *et al.*, 2009). Differences in V_f between individual and simultaneous uptake were compared using paired *t*-tests for each nutrient separately and then in unison. All data analyses, including regressions for nutrient uptake length calculations, were performed using SPSS version 19.0 (SPSS Inc., Chicago IL).

6.3 Results

6.3.1 Hydrological and environmental river characteristics

River discharge ranged from 5 to 240 l s^{-1} (Table 6.1) and was representative of the small rivers found in this area of Svalbard during summer (Blaen *et al.*, *in press*). There was no evidence of substantial lateral hillslope groundwater inflow in any river, as indicated by little longitudinal change in Cl^- concentration along the study reaches. Ambient NH_4^+ concentration was <20 μg l^{-1} at all sites and ambient acetate and PO_4^{3-} concentration was generally below detectable limits (<2 and <5 μg l^{-1}, respectively). Ambient NO_3^- concentration was more variable, ranging from 7 to 85 μg l^{-1} (Table 6.1). Ambient nutrient concentrations were not correlated with any measured environmental variables. Hydrochemically-informed end-member mixing analysis indicated that meltwater relative flow contribution accounted for between 0% and 74% of total river discharge in the selected study reaches (Table 6.1). Meltwater relative contribution was not significantly correlated with environmental variables. Correlation analyses of environmental variables showed significant relationships between water temperature and incoming short wave radiation ($p<0.05$), chlorophyll *a* concentration ($p<0.05$) and river discharge ($p<0.05$), and between SSC and Pfankuch Index ($p<0.01$; Table 6.2).

Table 6.1: *Descriptions, background nutrient concentrations and environmental characteristics of the study rivers*

Year	*Site*	*Discharge* $l\ s^{-1}$	*Meltwater* *%*	*Width* *m*	*pH*	*Temperature* *ºC*	*Suspended sediment* $mg\ l^{-1}$	*Conductivity* $\mu S\ cm^{-1}$	*Pfankuch Index*	NO_3^- $\mu g\ l^{-1}$	NH_4^+ $\mu g\ l^{-1}$	*Acetate* $\mu g\ l^{-1}$	PO_4^{3-} $\mu g\ l^{-1}$	*Chl a* $mg\ m^{-2}$	*Sediment respiration* $\mu g\ O2\ hr^{-1}\ kg^{-1}$
2011	X1	5	56	2.3	8.4	4.6	1	179	24	9	-	-	-	0.89	157
	X2	17	59	1.2	8.7	3.4	5	129	29	85	8	-	5	0.89	4
	X3	53	65	2.3	8.7	6.9	8	140	24	14	14	-	-	1.13	130
	X4	6	64	1.0	8.6	4.9	0	207	19	62	17	-	-	1.07	77
	X5	9	42	1.5	8.7	4.9	0	124	17	78	12	-	-	0.53	90
	X6	19	74	1.4	8.6	2.1	140	66	57	7	16	-	-	0.00	0
2012	Y1	240	45	2.6	7.8	11.4	1	143	34	21	5	-	-	2.69	40
	Y2	46	33	2.1	8.8	6.9	0	138	20	80	10	5	5	0.89	59
	Y3	187	15	2.8	8.9	5.8	0	66	20	20	10	-	-	2.97	86
	Y4	16	72	1.2	8.4	6.8	0	60	17	69	4	28	-	9.01	130
	Y5	14	52	1.0	9.0	5.1	0	78	15	70	16	20	-	7.95	94
	Y6	37	0	1.0	8.7	8.8	0	181	27	22	10	-	5	4.45	120

Table 6.2: *Spearman correlation coefficients for relationships between meltwater flow contributions and environmental habitat variables*

	% Meltwater	*Discharge*	*Pfankuch Index*	*EC*	*SW radiation*	*Water temperature*	*SSC*	*Chlorophyll a*	*As/A*
Discharge	-0.29								
Pfankuch Index	0.11	0.52							
EC	-0.27	-0.08	0.22						
SW radiation	-0.26	0.53	0.14	-0.02					
Water temperature	-0.42	0.62 *	-0.11	0.26	0.61 *				
SSC	0.57	0.27	0.76 **	-0.12	-0.20	-0.31			
Chlorophyll a	-0.21	0.25	0.23	-0.07	0.60	0.65 *	-0.51		
A_s/A	-0.12	-0.50	-0.28	-0.32	-0.38	-0.18	-.200	-0.22	
T_{stor}	-0.03	-0.20	-0.13	-0.67	-0.37	-0.14	.090	-0.42	0.90 **

** and ** denote $p<0.05$ and $p<0.01$, respectively.*   *$n=12$ except for A_s/A and T_{stor} where $n=8$*

Concentrations of NO_3^- and NH_4^+ in meltwater and groundwater samples were low (<100 µg l^{-1}) and there were no significant differences in nutrient concentrations between sample types for any of the studied rivers (p>0.05 by t-test). Acetate was normally below detectable limits in all samples. However, PO_4^{3-} concentration was significantly higher in groundwater samples (mean±SD: 88±15 µg l^{-1}) than in meltwater samples (mean±SD: 46±13 µg l^{-1}) at all sites (t=2.82-9.15, p<0.05). Mean water temperature over the injection periods ranged from 2.1 to 11.4 ºC. Chlorophyll *a* concentration ranged from 0 to 9.01 mg m^{-2} and rates of sediment respiration ranged from 0 to 157.1 µg O_2 h^{-1} kg^{-1} (Table 6.1). Transient storage area (As/A) ranged from 0.14 at site X4 to 3.08 at site Y2 (Table 6.3). Damkohler I numbers (DaI) were between 0.7 and 4.4 and within a range acceptable for reliable parameter estimation (Edwardson *et al.*, 2003). Mean storage zone residence times were related closely to transient storage (r=0.90, p<0.01) and ranged from <1 min to 22 min. Transient storage area was not related significantly to river discharge. The proportion of river discharge comprised of meltwater was not significantly correlated with ambient nutrient concentrations, measured environmental variables or transient storage parameters (Table 6.2; p>0.05).

Table 6.3: *Optimised output parameters and Damkohler I numbers from OTIS-P model simulations for the study rivers*

Site	*Relative storage zone area, A_s/A*	*Storage zone exchange coefficient, α* *s^{-1}*	*Mean storage zone residence time, T_{stor}* *min^{1}*	*DaI*
A1	1.0	0.0026	6.3	1.7
A2	-	-	-	-
A3	1.3	0.0014	15.5	0.7
A4	0.1	0.0005	4.4	2.6
A5	-	-	-	-
A6	-	-	-	-
B1	-	-	-	-
B2	3.1	0.0023	22.4	1.1
B3	0.9	0.0036	4.3	1.1
B4	0.5	0.0034	2.3	4.4
B5	3.0	0.0034	14.8	2.3
B6	0.8	0.0024	5.7	1.5

Sites where storage exchange parameters could not be estimated reliably are indicated by -

6.3.2 Nutrient uptake characteristics

At many sites there was no statistically significant downstream change in nutrient concentration after dilution corrections during each short-term release ($p>0.05$). In rivers where significant changes in concentration occurred, uptake length (S_w) for all nutrients was <130 m (Table 6.4). Only two sites (X3 and Y2) exhibited significant NO_3^- uptake, although NO_3^- uptake velocity (V_f) was generally higher than for other nutrients (Figure 6.3). Significant NH_4^+ uptake occurred at three sites (X1, X3 and Y2) when added singularly. However, when added in conjunction with acetate five sites displayed significant NH_4^+ uptake, where V_f ranged from 0.22 to 4.03 m h^{-1} and areal uptake rate (U) ranged from 0.07 to 175.67 µg m^{-2} h^{-1} (Table 6.3). Mean S_w for NH_4^+ shortened by 47 m when added in conjunction with acetate compared to as a single addition, and V_f and U both increased. Similarly, acetate uptake during individual injections was significant only at three sites, but this increased to five sites during the simultaneous injection with NH_4^+. Individual injection acetate S_w (41-47 m) was lower consistently than individual injection NH_4^+ S_w (64-123 m). Acetate uptake response to simultaneous injections was more variable than for NH_4^+ uptake with both increases, decreases, and no change in S_w observed between sites. The mean demand ratio for NH_4^+ was 3.2 (±1.2 SE; Figure 6.4), indicating an increase in demand for NH_4^+ when added in conjunction with acetate. In contrast, the mean demand ratio for acetate was 1.1 (±0.4 SE), indicating that the simultaneous addition of NH_4^+ had little influence on acetate uptake. However, paired t-tests showed no significant difference in V_f between individual and dual additions for both NH_4^+ ($t=1.14$, $p=0.37$) and acetate ($t=0.25$, $p=0.82$).

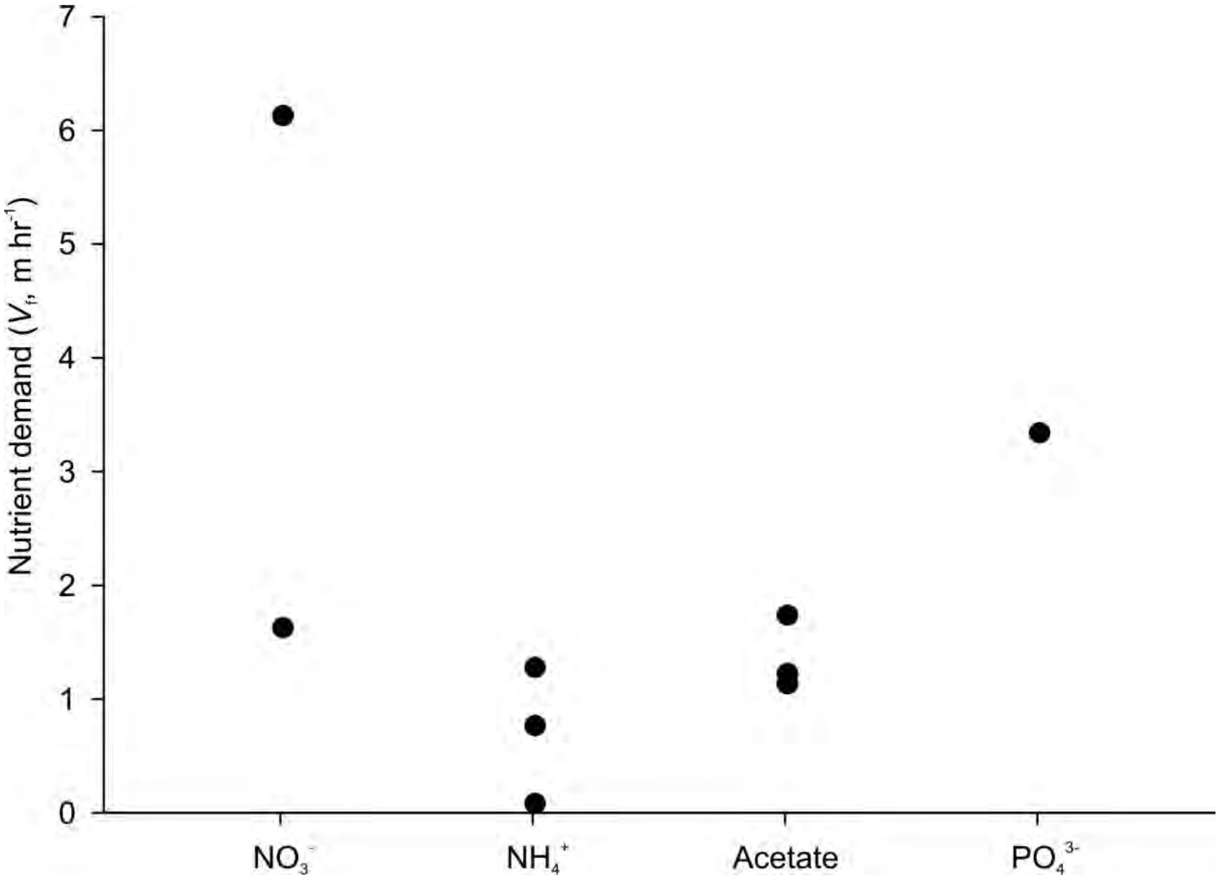


Figure 6.3: *Nutrient uptake velocity (V_f) for nitrate, ammonium, acetate and phosphate for single nutrient releases in the study rivers.*

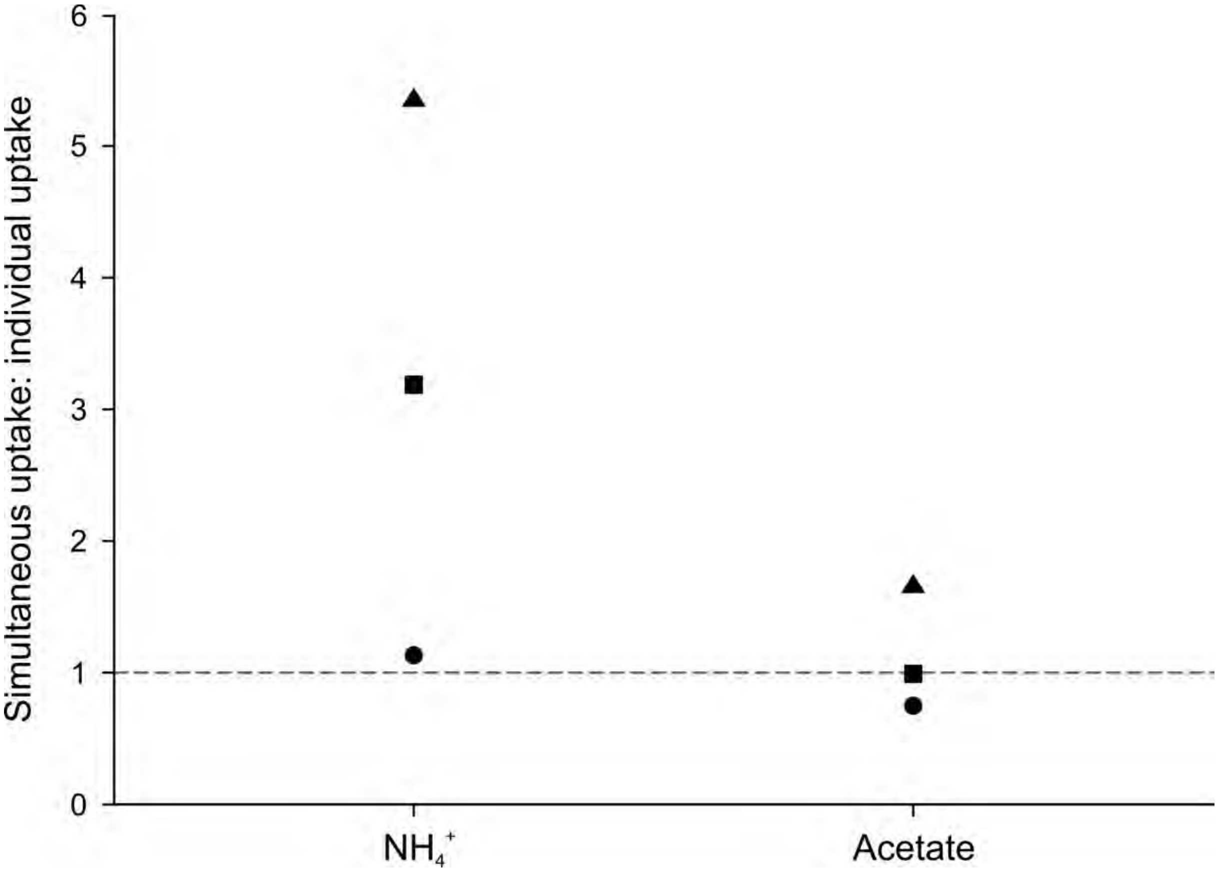


Figure 6.4: *Ratios of simultaneous to individual nutrient uptake velocity (V_f) for ammonium and acetate. Shapes indicate different streams. Nutrient uptake velocity was measured individually and then compared to when solutes were released together (i.e. $NH_4^+{}_{+acetate}\ V_f : NH_4^+\ V_f$)*

Table 6.4: *Nutrient uptake parameters for the study rivers. Uptake length (S_w) is reported in m, uptake velocity (V_f) in m hr^{-1}, and areal uptake rate (U) in µg m^{-2} hr^{-1}*

Addition		Site											
		X1	X2	X3	X4	X5	X6	Y1	Y2	Y3	Y4	Y5	Y6
NO_3^-	S_w	-	-	16.29	-	-	-	-	47.70	-	-	-	-
	V_f	-	-	6.12	-	-	-	-	1.62	-	-	-	-
	U	-	-	83.95	-	-	-	-	113.2	-	-	-	-
NH_4^+	S_w	123.5	-	64.94	-	-	-	-	102.3	-	-	-	-
	V_f	0.07		1.27	-	-	-	-	0.75	-	-	-	-
	U	0.02	-	17.18	-	-	-	-	7.54	-	-	-	-
NH_4^+ (+ Acetate)	S_w	80.65	90.39	51.40	50.51	-	-	-	19.13	-	-	13.35	-
	V_f	0.22	0.54	1.41	0.55	-	-	-	4.03	-	-	3.90	-
	U	0.07	4.11	19.20	9.15	-	-	-	72.02	-	-	175.7	-
Acetate	S_w	-	41.20	-	-	-	-	-	44.74	-	-	46.50	-
	V_f	-	1.21	-	-	-	-	-	1.72	-	-	1.12	-
	U	-	12.09	-	-	-	-	-	1.72	-	-	224.1	-
Acetate (+ NH_4^+)	S_w	-	40.10	-	-	-	-	65.20	61.13	-	25.25	28.24	-
	V_f	-	1.23	-	-	-	-	1.07	1.26	-	1.81	1.84	-
	U	-	12.56	-	-	-	-	1.08	1.98	-	506.8	369.0	-
PO_4^{3-}	S_w	-	-	-	-	-	-	-	-	-	-	15.60	-
	V_f	-	-	-	-	-	-	-	-	-	-	3.33	-
	U	-	-	-	-	-	-	-	-	-	-	3.79	-

- denotes no significant change in nutrient concentration throughout the study reach

Correlation analyses between NH_4^+ and acetate uptake parameters and environmental variables were conducted for dual additions only due to a limited number of cases (i.e. rivers with measurable uptake) for single nutrient injections. For NH_4^+, significant positive relationships were observed between V_f and water temperature (r=0.84, p<0.1), V_f and transient storage area (r=0.90, p<0.05), and U and chlorophyll *a* concentration (r=0.81, p<0.05) (Table 6.5). For acetate, V_f was negatively related to discharge (r=-0.90, p<0.05) and Pfankuch Index (r=-0.99, p<0.01), but positively related to sediment respiration (r=0.90, p<0.05). As with V_f, acetate, U was negatively related to Pfankuch Index (r=-0.90, p<0.05) and also to EC (r=-0.90, p<0.05). Variation in NH_4^+ and acetate V_f and U was not significantly explained by any other measured environmental variable, including ambient nutrient concentrations.

Table 6.5: *Significant Spearman correlation coefficients for nutrient uptake parameter relationships with environmental variables*

		n	*Discharge*	*Water temperature*	*Pfankuch Index*	*Electrical conductivity*	*Chlorophyll a*	*Sediment respiration*	A_s/A ^
NH_4^+ *(+ Acetate)*	V_f	6	0.49	0.84 *	-0.58	-0.43	0.64	-0.26	0.90 **
	U	6	0.37	0.71	-0.70	-0.54	0.81 **	-0.14	0.80
Acetate $(+ NH_4^+)$	V_f	5	-0.90 **	-0.40	-0.99 ***	-0.80	0.60	0.90 *	0.40
	U	5	-0.80	-0.30	-0.90 **	-0.90 **	0.70	0.90 *	0.20

, ** and * denote $p<0.1$, $p<0.05$ and $p<0.01$, respectively* *^ indicates $n=5$*

6.4 Discussion

This study offers new insights into potential changes in nutrient cycling in Arctic rivers that may occur with future climatic warming and changed water source partitioning in high-latitude regions. Demand for NH_4^+ was greater in warmer rivers with larger transient storage areas, while those with more algal biomass were associated with higher NH_4^+ uptake. Sediment respiration was related positively to acetate demand, suggesting that hyporheic zones may be important zones of carbon processing in these Arctic systems. Furthermore, the addition of a labile form of DOC stimulated NH_4^+ demand, indicating that labile DOC availability may limit some freshwater biotic process rates in this region of Svalbard. The following section discusses firstly the influence of changing hydrology on environmental conditions and nutrient uptake in Arctic river systems. The section then considers potential drivers of nutrient uptake and assesses the role of DOC availability in controlling NH_4^+ cycling.

6.4.1 Environmental habitat quality along a water source gradient

Water source partitioning is considered to exert a strong influence on environmental habitat conditions in many headwater basins in Arctic and alpine regions (Brown *et al.*, 2009; Milner *et al.*, 2009; Crossman *et al.*, 2011; Parker and Huryn, 2011). Rivers with glacial meltwater-dominated flow regimes are associated typically with very cold, turbid, unstable and nutrient-limited habitat conditions for aquatic organisms (Milner *et al.*, 2001; Huryn *et al.*, 2005). In contrast, conditions in rivers dominated by non-glacial sources (i.e. hillslope groundwater) are characterised generally by warmer temperature, greater solute enrichment, lower sediment loads and higher channel stability (Ward, 1999; Brown *et al.*, 2006b; Crossman *et al.*, 2011). The end-member mixing model used in this study showed that the twelve study sites spanned a water source gradient with meltwater comprising 0-74% of total river discharge, with the remaining fraction (26-100%) comprised of shallow hillslope groundwater. In Kongsfjorden, like most areas of Svalbard, deep groundwater-surface water interactions are thought to be limited by the presence of a widespread permafrost layer that acts as an aquitard to discharge of sub-permafrost water (Haldorsen *et al.*, 2002; Haldorsen *et al.*, 2010). Therefore, the groundwater flow component in the study rivers was assumed to be comprised predominantly of water that reached channels by subsurface flow through the shallow active layer (Wadham *et al.*, 2001). Interestingly, solute modelling indicated that the relative sizes of transient storage areas were comparable to, and some cases slightly higher than, those measured in Alaskan Arctic tundra systems and headwater rivers in temperate regions (Edwardson *et al.*, 2003; Zarnetske *et al.*, 2007; Scordo and Moore, 2009), suggesting that sub-channel permafrost had little effect on the size of transient storage zones in these high-latitude river systems.

The first hypothesis, that a decrease in meltwater contribution to total river flow would be associated with warmer and more stable channels that supported higher rates of autotrophy and microbial activity, was rejected because no significant relationships were observed between meltwater contribution and any of the measured environmental variables. However, the significant relationship between water temperature and chlorophyll *a* concentration suggested that warmer rivers were associated with greater algal growth. Low water temperatures observed in all study rivers (maximum: 11.4 ºC) reflected the reduced magnitude of atmospheric energy inputs at high latitudes, runoff interactions with cold permafrost, and glaciers and snowpacks as a primary source of river discharge (Irons and Oswood, 1992; Blaen *et al.*, *in press*). Proglacial lakes can influence river thermal regimes (Hieber *et al.*, 2002), but only one site (X1) was sourced from a lake outflow and water temperature at this location (4.6 ºC) was not markedly different from other sites (mean: 6.0 ºC). Strong associations between water temperature and incoming shortwave radiation suggested the latter was a principal driver of thermal and flow variability in these Arctic rivers. This relationship is often pronounced in high-latitude or mountainous environments above the treeline where little riparian shading occurs (Hannah *et al.*, 2008). These findings suggest that atmospheric energy receipt may be more important than water source contribution in influencing river temperature in this region of Svalbard (Blaen *et al.*, *in press*) and contrast with previous studies in alpine regions where relative meltwater contribution has been shown to have a strong influence on river habitat conditions (Brown *et al.*, 2007b; Brown *et al.*, 2010).

6.4.2 Nutrient uptake along a water source gradient

Most rivers showed no response to additions of NO_3^- and PO_4^{3-}, and for four rivers (X5, X6, Y3 and Y6) there were no significant downstream changes in concentration of any nutrients during the injection experiments. At these locations, river reaches appeared to act as neutral ‗pipes' through which nutrients were transported largely passively (Cole *et al.*, 2007), with no net transformation or uptake from the water column. Other studies have shown that a lack of uptake can be caused by saturation of biotic demand (Haggard *et al.*, 2001), although this is more common to temperate agricultural areas with high ambient nutrient concentrations. The extremely low ambient concentrations observed in these Svalbard study rivers (<0.1 mg l^{-1} at all sites for all nutrients), combined with low enrichment levels during each nutrient addition, suggests that saturation is unlikely to have occurred (Hoellein *et al.*, 2007). While chlorophyll *a* concentration at all sites was low relative to other Arctic, sub-Arctic and alpine rivers (Uehlinger *et al.*, 2010; Gudmundsdottir *et al.*, 2011; Parker and Huryn, 2011), site X5 was particularly deficient (0.53 mg m^{-2}) and no chlorophyll *a* was detected at site X6, suggesting that the algal capacity for nutrient uptake at these sites was almost nonexistent. Furthermore, the very cold, turbid and dilute water at site X6 was likely to have played a major role in regulating uptake in this channel. Flashy river flow regimes associated with Arctic meltwater systems (Hodgkins *et al.*, 2009) can scour bed sediments, leading to low channel stability and the erosion of biofilms responsible for nutrient uptake in these rivers (Aldridge *et al.*, 2010). However, uptake lengths for NO_3^- are longer typically than those of NH_4^+ (Newbold *et al.*, 2006) and it is possible that the 70 m reaches used in this study were not of sufficient length to measure a significant decline in NO_3^- concentration given the very harsh habitat conditions. Additionally, the short duration of the nutrient addition experiments (usually < 1 h) may not have been sufficiently long to stimulate biotic nutrient assimilation in some study rivers. Moreover, high dissolved oxygen concentrations in many study rivers (P. Blaen, unpublished data) may have inhibited rates of denitrification in bed sediments (Zarnetske *et al.*, 2012). Finally, while P availability can limit biological activity in some Arctic freshwater systems (Lock *et al.*, 1990; Säwström *et al.*, 2007; Stibal *et al.*, 2008), the lack of response to PO_4^{3-} injections indicates that this may not be the case for rivers in this region of Svalbard.

In rivers where significant uptake occurred, short uptake lengths (often <100 m) indicated high biotic demand and efficiency in retaining added nutrients (Dodds *et al.*, 2004), and highlighted the importance of headwater rivers in regulating nutrient supply to downstream reaches or the ocean (Alexander *et al.*, 2000; Balestrini *et al.*, 2012). In contrast to NO_3^-, significant NH_4^+ uptake was observed at three sites when added singularly and at six sites when added in conjunction with acetate. Due to the nature of the methodology employed, measured declines in nutrient concentration can only be quantified as gross retention and more detailed information on the specific mechanisms involved (e.g. biotic assimilation, volatilisation, nitrification) are absent (Ye *et al.*, 2012). However, it is suggested that the difference in response between different forms of N may be due to the lower energy costs associated with NH_4^+ uptake compared to those with NO_3^-, resulting in preferential uptake of this nutrient by algae and heterotrophic microbes (Marti and Sabater, 1996; Newbold *et al.*, 2006). The link between biotic assimilation and algal growth is supported by the significant positive relationship between chlorophyll *a* concentration and the rate of NH_4^+ uptake (Hall and Tank, 2003). Hodson *et al.* (2005) noted that NH_4^+ in proglacial channels on the Brøgger Peninsula was less conservative than other species of N and suggested that significant retention of this nutrient is a common feature of rivers in Svalbard. However, at the two sites (X3 and Y2) where significant NO_3^- uptake occurred, uptake velocity was greater than for both NH_4^+ and acetate, suggesting demand for NO_3^- in these rivers was also high.

The second hypothesis was that a decrease in the relative proportion of water contributing to river discharge would be associated with higher nutrient uptake velocities and rates of retention due to the conditions typically associated with groundwater-influenced aquatic habitats (Parker and Huryn, 2011; Crossman *et al.*, 2011). However, this hypothesis was rejected because no significant relationship was observed between meltwater flow contribution and nutrient uptake metrics, nor did meltwater contribution appear to influence physicochemical habitat properties (above). Interestingly, there was also no correlation between meltwater inflow and ambient river nutrient concentrations, and nutrient concentrations in meltwater end-member samples were not significantly different relative to those sourced from groundwater (except PO_4^{3-}). This was notable because previous studies have suggested that nutrient concentrations in Arctic rivers are enhanced frequently by leaching of water from soils in proglacial areas, although this is often most evident towards the end of the ablation season (i.e. late August) when soil nitrification rates are highest (Gersper *et al.*, 1980; Hodson *et al.*, 2002) and senescence of tundra vegetation reduces demand for nutrients (Tye and Heaton, 2008). This study was conducted primarily in mid-July when high demand for nutrients by plants may have resulted in rapid terrestrial uptake (Keuper *et al.*, 2012; Harms and Jones, 2012), thus reducing the flux of nutrients to river channels and possibly providing an explanation for the lack of significant differences in most nutrient concentrations between meltwater and groundwater end-member samples. Given that Arctic tundra ecosystems are often N-limited (Madan *et al.*, 2007; Kelley and Epstein, 2009), this may explain the surplus of PO_4^{3-} observed in groundwater end-member samples relative to meltwater samples. Alternatively, a secondary source may have introduced nutrients to rivers, possibly in the form of droppings from Barnacle Goose (*Branta leucopsis*) or seabird colonies common to Kongsfjord (Mindl *et al.*, 2007), which could mask a relationship between water source contributions and ambient river nutrient concentration.

6.4.3 Drivers of nutrient uptake in Arctic river systems

Low SSC at most sites and an absence of correlations between incoming short-wave radiation and uptake metrics suggested that light availability did not limit autotrophic activity in these rivers given the linear relationship between short-wave and photosynthetically active radiation (Howell *et al.*, 1983). However, uptake velocity for NH_4^+ was positively related to water temperature, suggesting that the low water temperature that characterises many rivers may act as a driving factor of ecosystem metabolism in this region of Svalbard (Brown *et al.*, 2004; Rasmussen *et al.*, 2011; Blaen *et al.*, *in press*). River metabolism is considered to be highly temperature-dependent (Acuña *et al.*, 2008; Demars *et al.*, 2011; Yvon-Durocher *et al.*, 2012) with warmer temperature associated with increased photosynthesis, enzymatic and microbial activity, and nutrient uptake (Barko and Smart, 1981; Rasmussen *et al.*, 2011).

Although areal-based uptake rates (U) were extremely low, demand (V_f) for NH_4^+ was comparable to demand observed previously in lower-latitude Arctic, sub-Arctic and temperate headwater rivers (Webster *et al.*, 2003; Ensign and Doyle, 2006; Rasmussen *et al.*, 2011). Unlike lower-latitude systems, rivers in Svalbard contain no macrophytes and primary production is dominated by algal communities (Kim *et al.*, 2008). The surface area available for nutrient uptake in algal communities is lower than in dense macrophyte stands (Rasmussen *et al.*, 2011) which is likely to limit the rate of nutrient uptake. However, the positive relationship between areal-based uptake rate for NH_4^+ and chlorophyll *a* concentration suggests that biotic assimilation by autotrophs played a substantial role in the removal of NH_4^+ from the water column, as opposed to other removal mechanisms such as nitrification or volatilisation (Webster *et al.*, 2003). Additionally, the positive relationship between NH_4^+ uptake velocity and transient storage zone area (A_s/A) indicates that increased hydraulic retention was associated with greater nutrient uptake capacity in these rivers. Conclusions of previous studies relating transient storage to nutrient uptake have been equivocal (Mulholland *et al.*, 1997; Hall *et al.*, 2002; Gucker and Boechat, 2004), but these results suggest temporary detainment of waters in side-pools or hyporheic zones increased residence times and the contact of solutes with sediment biofilms (Greenwald *et al.*, 2008), thus increasing the potential for biogeochemical transformation.

In contrast to NH_4^+, demand for acetate was not related significantly to chlorophyll *a* concentration because aquatic autotrophs obtain carbon predominantly through CO_2 and HCO_3^- (Allen and Spence, 1981; Cavalli *et al.*, 2012) whereas uptake of DOC is controlled to a large extent by heterotrophic activity (Marti *et al.*, 2009). The positive relationship between acetate demand and sediment respiration suggests that microbial communities in the hyporheic zone may play an important role in DOC processing in these Arctic rivers, as observed previously in alpine and lowland streams in lower-latitude areas (McKnight *et al.*, 1993; Newbold *et al.*, 2006; Johnson *et al.*, 2009). Similarly, the negative relationship between acetate demand and the Pfankuch Index, a measure of channel stability (Pfankuch, 1971), may be explained by more stable river channels promoting the growth and activity of heterotrophic organisms with a consequent rise in demand for DOC.

6.4.4 *Regulation of NH_4^+ uptake by DOC availability*

The third hypothesis was that the presence of a labile source of DOC would stimulate NH_4^+ uptake by releasing N assimilation from C limitation in heterotrophic microbial communities. Taken alone, significant acetate uptake at only three sites suggested low demand for labile DOC in the study rivers. However, the observed increase in demand for NH_4^+ when added in conjunction with acetate suggested that labile DOC availability does influence N uptake in these rivers. It was less clear whether DOC uptake was influenced by NH_4^+ availability: significant acetate uptake increased from three to five sites when added together with NH_4^+ but demand ratios showed no clear pattern in relation to those of NH_4^+. The limited number of cases meant that these differences in nutrient demand were not statistically significant between single and dual additions, although it should be noted that the test did not account for the rivers where significant uptake only occurred during the simultaneous additions. A previous study from Svalbard concluded that DOC availability regulated heterotrophic bacterial activity in lentic freshwaters in the Kongsfjorden area (Säwström *et al.*, 2007). Other studies have also established close links between the supply of DOC and demand for nitrogen in aquatic systems (Bernhardt and Likens, 2002; Dodds *et al.*, 2004; Johnson *et al.*, 2009) which is rooted in Redfield‘s (1958) seminal research on elemental stoichiometry. In light of these results the third hypothesis is cautiously accepted, but it is suggested that further research is now needed to substantiate these findings.

6.5 Conclusions and implications for nutrient cycling in Arctic rivers under a warming climate

This study investigated rates of nutrient uptake and links to controlling hydrological and environmental factors across a gradient of High Arctic river types. The proportion of river discharge derived from meltwater was not significantly related to measured environmental variables (Hypothesis I). A lack of significant removal of nutrients from the water column at several sites was attributed to cold water temperature, very low ambient nutrient concentrations, and low channel stability which together limited algal and heterotrophic microbial growth and thus the capacity for nutrient uptake. Where uptake did occur, short uptake lengths and high uptake velocities indicated high demand for nutrients in these systems. Uptake velocities for NH_4^+ were comparable to those observed in lower-latitude rivers, although areal-based uptake rates were low and limited probably by low algal biomass. Uptake of NH_4^+ was associated with warmer water temperature, algal growth and larger transient storage zones, while acetate uptake was associated with sediment respiration and more stable river channels. The proportion of river discharge comprised of meltwater was not related to nutrient uptake parameters (Hypothesis II). Some evidence was observed to suggest that NH_4^+ assimilation by heterotrophic microbial communities is mediated by labile DOC availability in these rivers (Hypothesis III), although a limited number of cases means that further research is required to explore more fully this potential relationship.

In the context of future warming in Arctic regions, shrinking glaciers, decreasing meltwater contributions and increased permafrost wasting are expected to result in a progressive shift towards more groundwater-dominated river systems (Smith *et al.*, 2007; Walvoord and Streigl, 2007; St. Jacques and Sauchyn, 2009). The prevailing view is that such a shift will be associated with a general increase in nutrient availability for freshwater biota in high-latitude riverine environments (Frey and McClelland *et al.*, 2009; Rowland *et al.*, 2010). Moreover, there is also growing concern over episodic atmospheric N deposition in the High Arctic from Western European pollution transported in the troposphere (Hodson *et al.*, 2010; Kühnel *et al.*, 2011). Together, these factors could introduce substantial variation in nutrient fluxes and availability into these historically sensitive and nutrient-limited environments (Shaver and Chapin, 1980). A combination of warmer water temperature, more stable river channels, greater thaw depths and increased nutrient loading is likely to increase autotrophic and heterotrophic activity in freshwater communities. The results of this study suggest that this may result in enhanced N retention in headwater Arctic rivers. A concurrent increase in C supply linked to permafrost degradation as warming continues (Vavrus *et al.*, 2012) may enhance the capacity of these systems to process nutrients. Ultimately, future nutrient yields from high-latitude river systems will be determined by changes in nutrient loading due to permafrost wasting and an increase in vegetation cover versus the degree to which aquatic autotrophs and heterotrophs can assimilate and process these nutrients in a changed riverine environment.

6.6 Chapter summary

An experimental approach was adopted to quantify water source contributions (meltwater, groundwater), environmental conditions and uptake of NO_3^-, NH_4^+, PO_4^{3-} and acetate at twelve headwater rivers in Svalbard and so determine the role of changing hydrology on nutrient uptake. The results emphasise the heterogeneity in nutrient uptake capacity among high-latitude rivers and indicate that future nutrient cycling in Arctic basins may be controlled by both physical and biological process interactions. Hydrological controls on freshwater ecology are explored further in Chapter 7 by examining how future environmental change might affect benthic macroinvertebrate communities in high-latitude rivers.

CHAPTER 7

HYDROLOGICAL DRIVERS OF MACROINVERTEBRATE COMMUNITIES IN ARCTIC RIVERS

7.1 Introduction

Hydrological change in high-latitude regions as a consequence of shifting water source contributions could cause large changes in physicochemical habitat conditions by altering thermal, nutrient, sediment and disturbance regimes in Arctic rivers (Judd and Kling, 2002; Holmes *et al.*, 2008; Milner *et al.*, 2009). In turn, these habitat changes are likely to affect the composition of riverine communities and could have substantial implications for aquatic biodiversity in high-latitude regions (Prowse *et al.*, 2006; Wrona *et al.*, 2006; Vincent *et al.*, 2011). Previous studies in Arctic rivers have implied that water source proportioning plays an important role in determining aquatic ecological community structure and function (Ward, 1994; Huryn *et al.*, 2005; Friberg *et al.*, 2013). Consequently, the quantification of water source-habitat-biota relationships could lead to the development of predictive models with which to assess potential changes in high-latitude riverine biodiversity in response to shifts in water source dynamics (*cf* Brown *et al.*, 2009).

The literature review (Chapter 2) illustrated that at present there is minimal understanding of how future hydrological changes in Arctic river basins will impact aquatic ecological communities. This research gap is important because freshwater taxa in these systems may be useful and sensitive indicators of high-latitude environmental change. Therefore, the aim of this chapter is to quantify changes in conceptual water source contributions to flow, habitat conditions, and benthic macroinvertebrate communities across several river basins of varying glacierization to infer how altered hydrology and associated changes in environmental habitat conditions may affect these Arctic river ecosystems in future. The following hypotheses were tested:

(I) decreased meltwater contributions will result in increased water temperature, EC and channel stability and lower SSC;

(II) decreased meltwater contributions will result in increased abundance and diversity of benthic macroinvertebrate communities, and;

(III) water source contribution optima can be identified for individual macroinvertebrate taxa.

7.2 Methodology

7.2.1 *Study area, site selection and sampling framework*

The study was conducted undertaken from 6 July to 30 August 2010 and from 5 July to 30 August 2011 near the Ny Ålesund research station in north-west Spitsbergen. Rivers in the area are sourced primarily from a combination of glacier and snow meltwater runoff and hillslope groundwater, the relative proportions of which vary on a seasonal and interannual basis (Chapter 4).

Seven sites were selected to characterise water source dynamics, river habitat condition, and ecological community dynamics. Three sites in the Bayelva basin were selected to provide a range of glacial influence over which to evaluate hydrological conditions near the glacier snout (V1), along the river mainstem (V2), and at the Bayelva river mouth (BR). Additional sites in Stuphallet cliffs area (S3; 2010 only) and the wider Bayelva area (N1, N2, N3; 2011 only) were identified to compare between glacierized and non-glacierized systems (Figure 7.1). Access to site S3 was by boat in 2010; however, this transportation option was not available in 2011. Consequently, sampling efforts were directed closer to Ny Ålesund in this year. Non-glacierized sites were considered to be representative of snowmelt and hillslope groundwater dominated river systems found in the Kongsfjord region of Svalbard.

Benthic macroinvertebrate samples collected from river sites were combined with river habitat characteristics and quantitative estimates of water source contributions to bulk discharge to understand hydrological controls on macroinvertebrate community dynamics in these Arctic river systems.

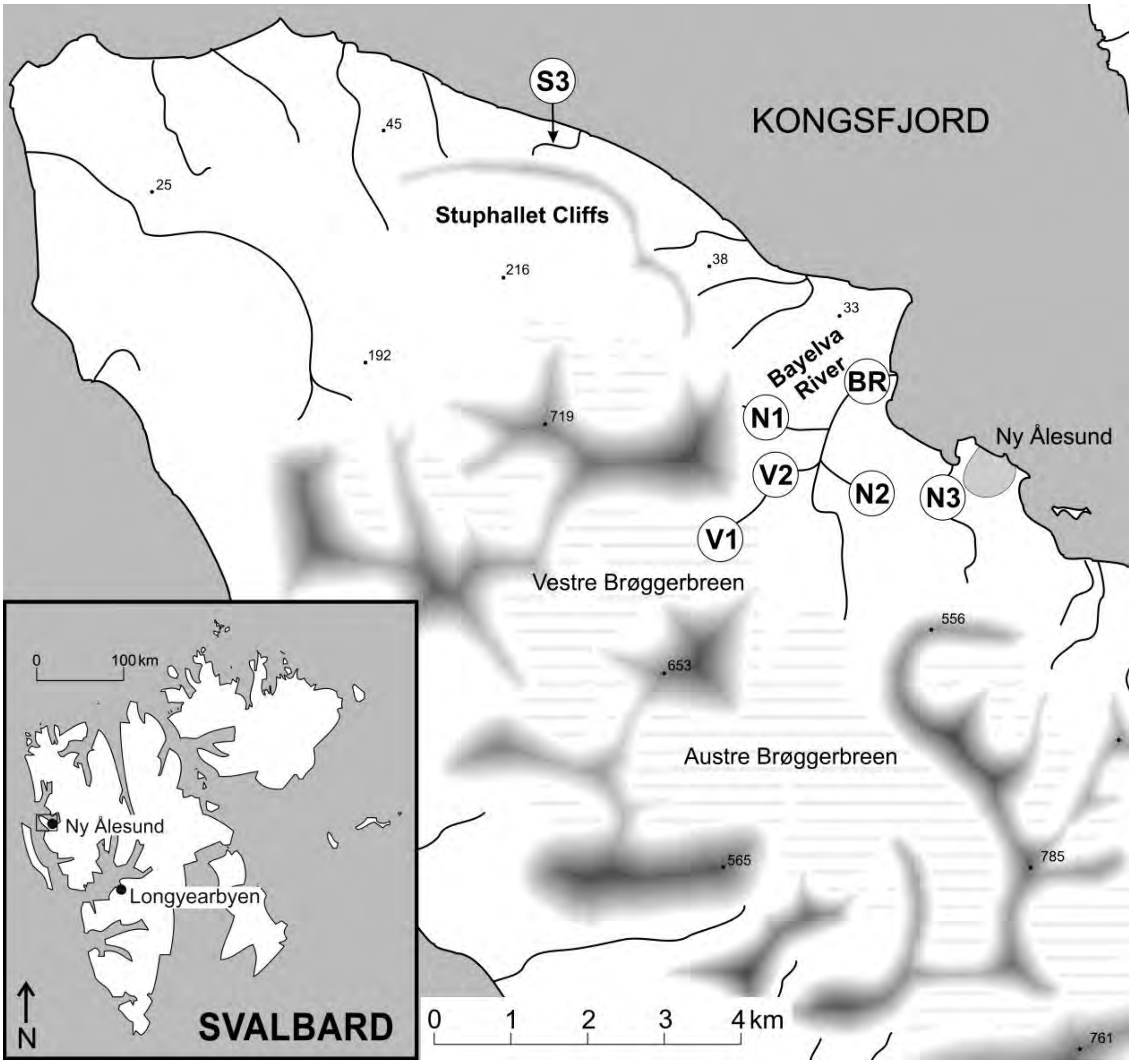


Figure 7.1: *Map of study area showing sampling sites, approximate river courses, relief (shaded), spot height measurements in meters asl, and glacier cover (dashed areas).*

7.2.2 Water source dynamics

Meltwater contributions to bulk river flow at each site were derived from hydrochemically-based end-member mixing analysis following the methodology outlined in Chapter 4. Separate estimates were produced for different sampling dates. Results are expressed as a proportion of total river flow.

7.2.3 Benthic macroinvertebrates and river habitat variables

Five replicate samples of benthic macroinvertebrates were collected on each sampling date. River discharge, water temperature, electrical conductivity (EC), suspended sediment concentration (SSC), pH, Pfankuch Index, bed sediment size and chlorophyll *a* concentration were determined immediately after macroinvertebrate sampling. Sampling and identification methods are provided in Chapter 3.

7.2.4 Data analysis

Benthic macroinvertebrate samples were collected from river sites every two weeks in 2010. In 2011, samples were collected every four weeks due to logistical constraints. Replicate macroinvertebrate samples were pooled to reduce the effects of path-scale variability on inter-reach comparisons of community structure (Brown *et al.*, 2007b). Pooled data were used to determine alpha diversity (taxonomic richness), beta diversity (Whittaker, 1972; average taxonomic richness from temporal replicate samples divided by pooled sample taxonomic richness, where lower values indicate higher beta diversity), total abundance, Shannon diversity index, and evenness.

Relationships between water source dynamics, habitat variables and macroinvertebrate abundance and diversity metrics were explored using generalized estimating equations (GEE). These models explicitly allow for autocorrelation between data acquired from the same site on multiple sampling occasions through the incorporation of a compound correlation structure (Hardin and Hilbe, 2003; Zuur *et al.*, 2010). Poisson distributions were used to facilitate analysis on response variables that were discrete or not normally distributed (Ballinger, 2004). Variance inflation factors (VIF) were used to assess colinearity between each explanatory variable used in the models. Variables with the highest VIFs were removed iteratively until all VIFs were <3 (Zuur *et al.*, 2010). Optimum model selection was conducted by removing explanatory variables and applying an analysis of variance (ANOVA) test to the model outcome. Model evaluation was performed by calculating a marginal R^2 value which represents the variance in the response variable explained by the fitted model (Ballinger, 2004). Validation of models was performed by examining residual plots for homogeneity and independence (Hardin and Hilbe, 2003; Zuur *et al.*, 2010). Non-metric multidimensional scaling (NMDS) was used to examine relationships between water source contribution, river habitat variables and macroinvertebrate community structure. NMDS was performed based on Bray-Curtis dissimilarities using $\log_{10}(x+1)$ abundance data with 100 random starts to avoid the risk of local optima (Soininen, 2004). Environmental variables were $\log_{10}$ transformed and fitted as vectors in the ordination space (1000 permutations). Variables correlated significantly with the ordination were added to the output plot. Meltwater optima (proportion of meltwater at which abundance peaks) and tolerance (one standard deviation around the optimum) for macroinvertebrate taxa were calculated using a weighted average transfer function with inverse deshrinking and bootstrap cross-validation (100 cycles). Water temperature optima were also calculated because water temperature is believed to be a key driver of benthic macroinvertebrate community dynamics in some Arctic rivers (Lods-Crozet *et al.*, 2001; Friberg *et al.*, 2013). Statistical tests were conducted using SPSS version 19.0 (SPSS Inc., Chicago IL). GEE modelling was performed using the *geepack* package (Halekoh *et al.*, 2006) and ordinations were performed using the *Vegan* package for R (Oksanen *et al.*, 2007), with significance set at $p<0.05$. Weighted average transfer functions were calculated using C^2 version 1.7.2 (Juggins, 2011).

7.3 Results

7.3.1 Water source contributions and river habitat variables: characteristics and relationships

Hydrochemically-informed end-member mixing analysis indicated that mean meltwater contributions from the shallow active layer ranged from 89% at site V1 (2010) to 29% at sites N1 and N3 (2011) of total river discharge in the selected study reaches (Table 7.1). Meltwater contributions to sites V1, V2 and BR in the glacierized Bayelva basin were high, particularly during 2010, and decreased marginally downstream with distance from the glacier. In non-glacierized basins, rivers were sourced to a lesser extent from meltwater contributions which typically comprised <60% mean total flow (Figure 7.2a). Discharge in glacier-fed rivers was 1-2 orders of magnitude greater than flows in non-glacierized systems (Table 7.1; Figure 7.2b). SSC was below detectable limits in non-glacier-fed rivers (Figure 7.2c). Water temperature, EC and channel stability were higher in non-glacierized river basins, and increased with distance from source in glacier-fed rivers (Figure 7.2d-7.2e; Table 7.1), but pH was higher in glacier-fed rivers (Figure 7.2f). In the Bayelva river basin, mean chlorophyll *a* concentration was greatest near the Vestre Brøggerbreen glacier (site V1) and decreased longitudinally downstream in both years (Figure 7.2g). Chlorophyll *a* concentrations in non-glacierized river basins showed greater variability, ranging from 71 mg m^{-2} at site S3 in 2010 to 2 mg m^{-2} at site N2 in 2011. GEE models showed meltwater contributions were significantly negatively related to EC and positively related to discharge, pH and Pfankuch Index (Marginal R^2: 0.14-0.45, $p<0.05$; Table 7.2).

Table 7.1: *Mean values (±1 standard deviation) of river habitat variables at monitoring sites.*

	n	*Meltwater proportion*	*Discharge* $m^3\ s^{-1}$	*Water temperature* ºC	*SSC* $g\ l^{-1}$	*EC* $\mu S\ cm^{-1}$	*Pfankuch Index*	*pH*	*Bed sediment* d_{50} *mm*	*Chlorophyll a* $mg\ m^{-2}$
2010										
V1	5	0.89 (0.08)	0.77 (0.68)	1.16 (0.54)	0.08 (0.07)	47.93 (14.44)	51	9.05 (0.16)	30	24.37 (7.00)
V2	5	0.86 (0.11)	0.94 (0.84)	4.32 (1.37)	0.10 (0.08)	62.02 (11.76)	34	8.68 (0.20)	8	18.12 (18.90)
BR	5	0.84 (0.14)	1.63 (1.46)	3.73 (1.14)	0.18 (0.05)	71.74 (34.18)	48	8.59 (0.15)	20	4.25 (0.29)
S3	4	0.62 (0.31)	0.03 (0.01)	3.22 (1.89)	0.00 (0)	172.14 (32.57)	16	8.60 (0.18)	5	71.43 (23.56)
2011										
V1	3	0.78 (0.29)	0.78 (0.21)	0.66 (0.39)	0.19 (0.22)	41.03 (7.33)	52	9.06 (0.23)	21	22.15 (13.74)
V2	3	0.77 (0.06)	1.54 (0.78)	3.40 (0.31)	0.12 (0.13)	55.86 (10.92)	34	8.82 (0.31)	10	19.58 (10.49)
BR	3	0.80 (0.06)	2.98 (1.15)	4.22 (0.86)	0.14 (0.13)	51.05 (13.46)	45	8.59 (0.39)	17	13.25 (3.26)
N1	3	0.29 (0.27)	0.04 (0.01)	4.91 (1.08)	0.00 (0)	137.42 (46.36)	22	8.43 (0.26)	10	8.11 (7.32)
N2	3	0.52 (0.27)	0.01 (0.01)	7.22 (1.26)	0.00 (0)	176.66 (132.71)	15	8.47 (0.09)	8	2.37 (2.06)
N3	3	0.29 (0.25)	0.07 (0.04)	5.61 (0.75)	0.00 (0)	219.28 (106.56)	30	8.45 (0.16)	10	6.13 (4.76)

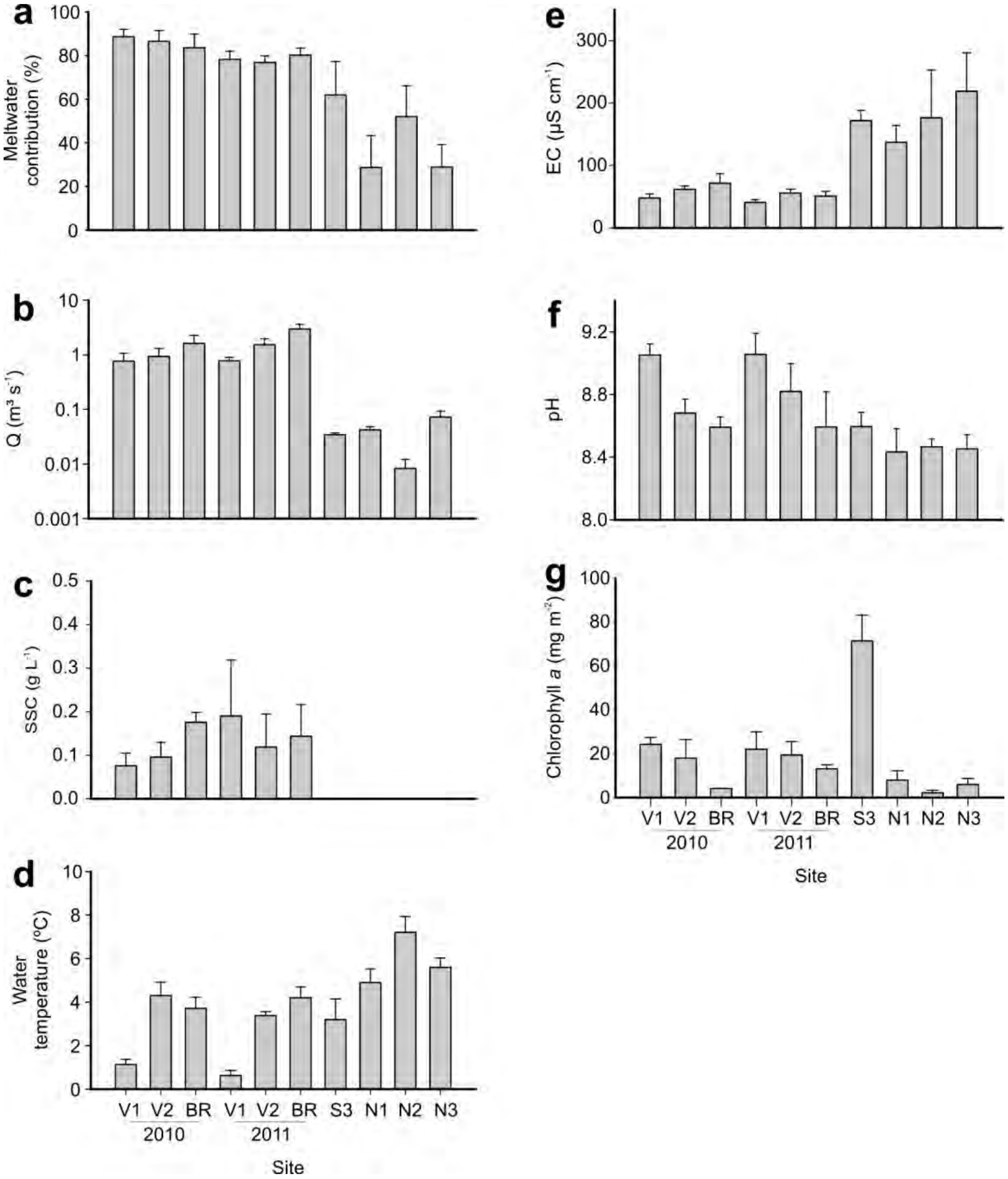


Figure 7.2: *Bar charts showing a) proportions of meltwater contributing to bulk river flow, b) discharge, c) SSC, d) water temperature, e) EC, f) pH and g) chlorophyll a concentration. Error bars show SE. n=5 (3) for sites V1, V2 and BR in 2010 (2011), n=4 for site S3 and n=3 for sites N1, N2 and N3*

Table 7.2: *Generalized estimating equation (GEE) model summary of relationships between meltwater contribution and river habitat variables. Standard errors are shown in parentheses.*

Response variable	*Intercept*	*Meltwater contribution*	*Marginal R²*
EC	5.523 (0.227) ***	-0.015 (0.004) ***	0.45
Pfankuch Index	3.540 (0.121) ***	0.0009 (0.00002) ***	0.28
Discharge	-4.344 (1.080) ***	0.054 (0.012) ***	0.24
pH	2.141 (0.005) ***	0.0002 (0.0001) *	0.14

****, ** and * denote significance at 0.001, 0.01 and 0.05 levels, respectively*

7.3.2 Influence of water source contribution and river habitat conditions on benthic macroinvertebrate communities

Chironomidae dominated benthic macroinvertebrate community structure at all sites and mean abundance ranged from 446 individuals m^{-2} (site N2) to 1558 individuals m^{-2} (site N1) (Table 7.3). Diamesinae were the major chironomid sub-family at all sites. Orthocladiinae were absent from site V1 at the snout of the Vestre Brøggerbreen glacier but their abundance increased both downstream and in rivers in non-glacierized basins (Table 7.3). Oligochaeta and Collembola typically comprised <10% of total mean abundance. Overall diversity was low and ranged from 1 at site V1 to 16 at site N2 (Figure 7.3a). Mean taxonomic richness, beta diversity, Shannon diversity and evenness increased longitudinally downstream from the glacier snout and were higher in most non-glacier-fed rivers (Figure 7.3a, 7.3c, 7.3d and 7.3e), but total abundance showed no clear pattern between sites (Figure 7.3b)

Table 7.3: *Mean abundance (±1 standard deviation) of macroinvertebrate taxa (per m^2) at monitoring sites.*

	Diamesinae	*Othocladiinae*	*Oligochaeta*	*Collembola*
2010				
V1	513 (782)	0 (0)	2 (4)	0 (0)
V2	716 (422)	121 (186)	25 (29)	8 (14)
BR	423 (429)	43 (41)	32 (10)	9 (9)
S3	1147 (879)	393 (421)	55 (27)	1 (1)
2011				
V1	993 (1103)	0 (0)	0 (0)	0 (0)
V2	978 (467)	573 (971)	171 (133)	2 (2)
BR	409 (496)	51 (56)	24 (10)	5 (1)
N1	1193 (1407)	366 (326)	123 (25)	144 (199)
N2	259 (211)	187 (154)	64 (32)	151 (176)
N3	401 (371)	73 (108)	201 (97)	41 (30)

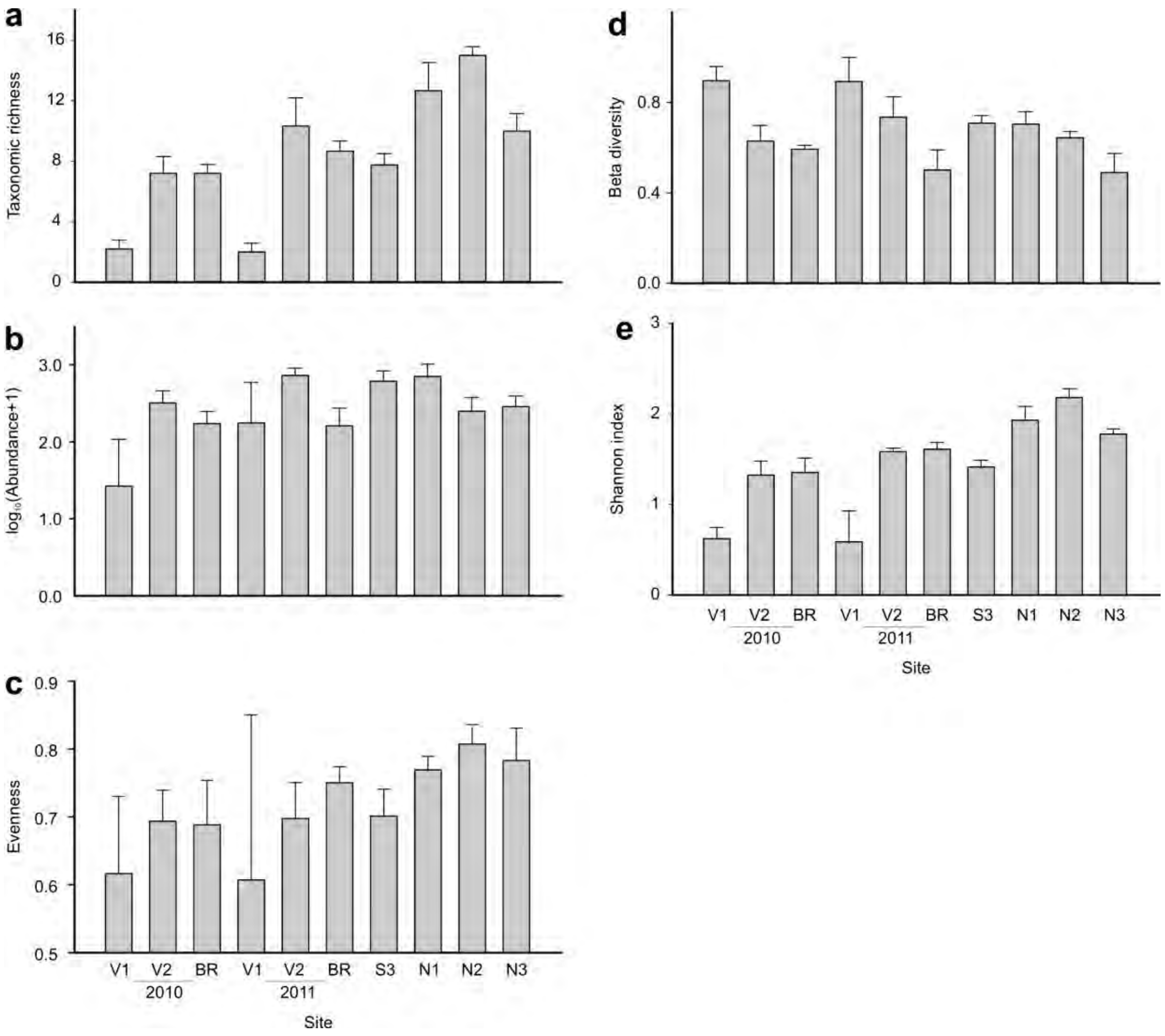


Figure 7.3: *Bar charts showing a) Taxonomic richness, b) Abundance, c) Evenness, d) Beta diversity and e) Shannon index. Abundance is shown log*$_{10}$*(x+1) transformed for clarity. Error bars show SE. n=5 (3) for sites V1, V2 and BR in 2010 (2011), n=4 for site S3 and n=3 for sites N1, N2 and N3.*

GEE models showed alpha diversity was significantly positively related to water temperature and negatively related to pH, bed sediment d_{50} and $\log_{10}$ chlorophyll *a* (Marginal R^2=0.72, p<0.05; Table 7.4). Total abundance was significantly negatively related to bed sediment d_{50} and Shannon diversity was significantly positively related to water temperature, pH and bed sediment d_{50} (Table 7.4).

A two-dimensional NMDS solution (stress=0.135) was chosen because additional dimensions resulted in only minor changes in stress (Soininen, 2004). Discharge was the only environmental variable significantly correlated with the ordination (r=0.41, p<0.05). Meltwater contribution was not included, although this variable exhibited covariance with discharge (r=0.49, p<0.05). Many species showed a preference for higher discharge (Figure 7.4). *Diamesa cinerella* gr./ *zernyi* gr. were located towards the negative (‗low discharge') end of axis 1 of the NMDS, whereas other Diamesinae and most Orthocladiinae taxa clustered towards the positive (‗high discharge') end.

Table 7.4: *Generalized estimating equation (GEE) summary of relationships between river habitat variables and macroinvertebrate community metrics. Standard errors are shown in parentheses.*

Response variable	*Intercept*	*Water temperature*	*pH*	*d50*	*Log $_{10}$ Chlorophyll a*	*Marginal R^2*
Alpha diversity	6.195 (1.365) ***	0.081 (0.213) ***	-0.443 (0.158) **	-0.036 (0.009) ***	-0.169 (0.041) *	0.71
Abundance	6.610 (0.215) ***			-0.042 (0.011) ***		0.14
Beta diversity	NS					
Shannon index	6.574 (1.617) ***	0.113 (0.014) ***	-0.609 (0.197) **	-0.024 (0.008) **		0.75
Evenness	NS					

****, *** and * denote significance at 0.001, 0.01 and 0.05 levels, respectively*

NS indicates no significant model was found

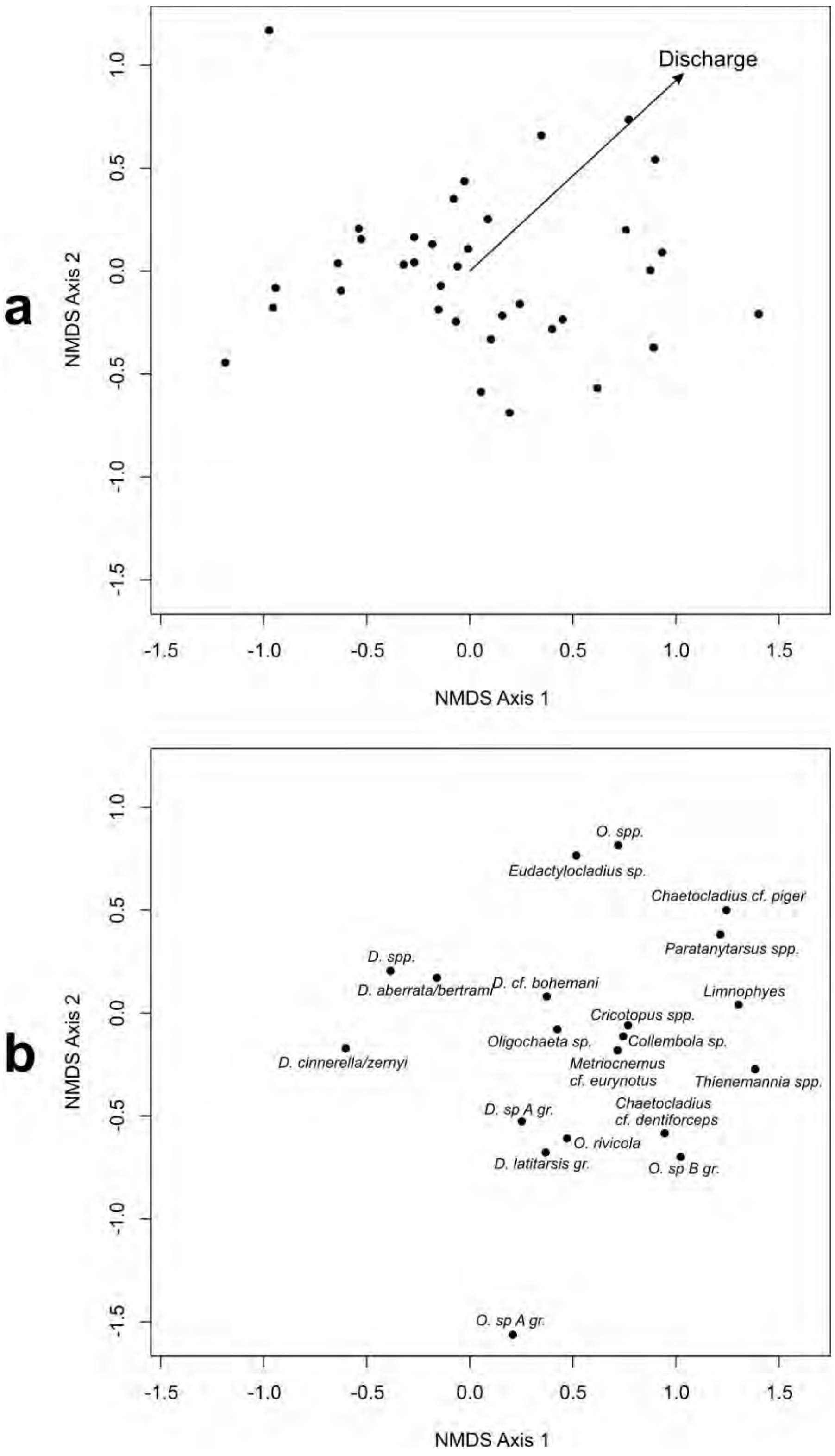


Figure 7.4: *Non-metric multidimensional scaling (NMDS) ordination plots of a) sites and significant environmental variables, and b) macroinvertebrate taxa. „O"and „D"denote Orthocladius and Diamesa, respectively.*

7.3.3 Water source contribution and temperature optima for individual macroinvertebrate taxa

Meltwater flow proportion optima for individual taxa ranged from 0.8 ±0.2 (*Orthocladius eudactcladius*) to 0.3 ±0.2 (*Diamesa latitarsis* gr., *Metriocnemus cf. eurynotus* and Collembola) (Table 7.5). There was no difference in mean meltwater optima between Diamesinae and Orthocladiinae taxa. Water temperature optima were also defined clearly: *D. cinerella* gr./*zernyi* gr. exhibited the lowest water temperature optimum (2.6 ±2.0 ºC) and *Orthocladius sp B* gr. and *Paratanytarsus* spp. the highest (7.5 ±1.4 and 6.4 ±1.1 ºC, respectively) (Table 7.5). The mean water temperature optimum for Diamesinae was 0.7 ºC lower than for Orthocladiinae. Negative correlations were observed between meltwater contribution optima and tolerance (r=-0.47, p=0.05) and water temperature optima and tolerance (r=-0.45, p=0.05).

Table 7.5: *Meltwater contribution and water temperature optima and tolerance values for macroinvertebrate taxa. Results are ordered by meltwater contribution optima.*

	Meltwater (proportion)		*Water temperature (°C)*	
	Optimum	*Tolerance*	*Optimum*	*Tolerance*
O. eudactcladius	0.84	0.16	4.41	1.19
Orthocladiinae spp.	0.83	0.12	4.39	1.09
D. cinerella gr./*zernyi* gr.	0.71	0.20	2.59	2.02
D. cf. bohemani	0.67	0.24	5.01	1.32
D. aberatta gr./*bertrami* gr.	0.67	0.25	4.12	1.99
Orthocladius sp A gr.	0.66	0.23	3.68	1.62
Diamesa spp.	0.64	0.25	3.79	2.20
Cricotopus spp.	0.60	0.26	4.63	1.47
O. rivicola	0.58	0.20	4.58	1.59
C. piger	0.55	0.22	5.87	0.75
Limnophyes spp.	0.55	0.17	4.80	1.19
Thienemannia spp.	0.53	0.17	4.45	1.01
Paratanytarsus spp.	0.51	0.14	6.38	1.14
Orthocladius sp B gr.	0.50	0.25	7.50	1.41
Oligochaeta	0.49	0.26	4.67	1.66
D. sp A gr.	0.45	0.26	4.92	1.50
Chaetocladius dentiforceps	0.39	0.23	5.04	1.45
Metriocnemus cf. eurynotus	0.35	0.20	4.78	2.08
Collembola	0.34	0.28	6.10	1.65
D. latitarsis gr.	0.33	0.26	5.42	1.31

7.4 Discussion

This research aimed to quantify relationships between water source contributions, river habitat variables, and benthic macroinvertebrate communities in Arctic rivers. Significant relationships were observed between meltwater contribution and several river habitat variables, suggesting that water source influenced physicochemical habitat quality in these river systems. Of these variables, however, only pH was related to benthic macroinvertebrate community metrics and GEE results showed water temperature and bed sediment size were better predictors of community metrics. A weighted average transfer function identified clear water source and water temperature preferences for most taxa, and also suggested that taxa associated with high meltwater contributions and cold water temperature may be resilient to climate warming.

7.4.1 Relationships between water source contribution and river habitat variables

The strong influence of water source (i.e. glacier runoff, snowmelt, groundwater) on environmental conditions in headwater rivers is well documented (Ward, 1994; Malard *et al.*, 2000; Brown *et al.*, 2003; Ilg and Castella, 2006; Parker and Huryn, 2011), but it is only recently that relationships have been drawn between quantitative estimates of water source contribution and aquatic habitat variables (Brown *et al.*, 2007b; 2009). Meltwater-dominated flow regimes can result in highly unstable habitats due to high variability in runoff and sediment loads (Wrona *et al.*, 2005). In contrast, groundwater-fed rivers are often characterised by less variable flow regimes (Prowse *et al.*, 2006; Blaen *et al.*, *in press*). In Svalbard, significant relationships between meltwater contribution and several habitat variables supported the first hypothesis of the study by indicating water source contribution played a major role in influencing habitat conditions in these Arctic river systems. Pfankuch Index decreased with reduced meltwater contribution and SSC was negligible in non-glacier-fed rivers, suggesting that rivers sourced largely from groundwater create less disturbed habitat conditions than rivers dominated by meltwater (Milner *et al.*, 2001; Parker and Huryn, 2011). Increases in EC and decreases in pH associated with reduced meltwater contributions support this hypothesis because ionic enrichment most likely reflects solute acquisition from active layer soils during subsurface flow (Cooper *et al.*, 2002; Hodson *et al.*, 2002), while declines in pH are associated with soil development in proglacial areas (Anderson *et al.*, 2000).

Warmer water temperature in headwater streams in high-latitude or mountainous areas has often been considered synonymous with reduced glacial influence and increased flow contributions from non-glacial water sources (Ilg and Castella, 2006; Brown *et al.*, 2007b; Füreder, 2007). However, in this study no significant relationship was observed between water source contribution and water temperature. River thermal regimes in Svalbard are often closely related to incoming solar radiation (Blaen *et al.*, *in press*). Therefore, these results suggest that the influence of water source contribution on water temperature in this region of Svalbard is not as great as in rivers in lower-latitude regions.

Some of the highest concentrations of chlorophyll *a* were recorded at site V1 in close proximity to the Vestre Brøggerbreen glacier with subsequent decreases in concentration downstream. Most previous studies in Arctic and alpine rivers have documented minimal algal development near glacier snouts due to high turbidity, shear stress and unstable substrate (Uehlinger *et al.*, 1998, Lods-Crozet *et al.*, 2001). Similar physical conditions were observed in this study: the lowest channel stability and highest bed sediment size were recorded in both years at site V1, although SSC increased longitudinally downstream because of sediment delivery from the proglacial sandur (Hodson *et al.*, 1998). Consequently, it is possible that high algal growth at this site may arise due to relatively high width:depth ratios near the glacier snout resulting in greater light availability than further downstream (supported by field observations), combined with a low but consistent supply of dissolved inorganic nutrients from glacial drainage (Hodson *et al.*, 2005).

7.4.2 Benthic macroinvertebrate community relationships with water source contributions and habitat variables

Water source contribution was not a significant predictor of macroinvertebrate community metrics, but sites in non-glacierized basins with a low proportion of meltwater had both greater taxonomic richness and abundance than those in glacierized river basins. Previous studies of river ecosystems in Svalbard have also suggested chironomid communities are more diverse in non-glacier-fed systems (Lods-Crozet *et al.*, 2001; Lods-Crozet *et al.*, 2007; Marziali *et al.*, 2009). Importantly, several chironomid taxa were found only at sites in non-glacierized river basins (e.g. *Metriocnemus* cf. *eurynotus*, *Paratanytarsus* spp.). These taxa are associated with more stable habitat conditions (Armitage *et al.*, 1995), indicating that non-glacier-fed river systems may play an important role in supporting regional macroinvertebrate biodiversity in High Arctic areas. Similar observations have also been made in some alpine and Arctic Alaskan rivers where spring-fed systems support greater macroinvertebrate taxon richness and abundance than glacier-fed rivers (Ilg *et al.*, 2001; Parker and Huryn, 2011). In Svalbard, this finding most likely reflects the less harsh physico-chemical habitat conditions associated with reduced meltwater contributions to rivers in non-glacierized basins (Füreder *et al.*, 2001; Brown *et al.*, 2007b), thus leading to partial support for the study‘s second hypothesis.

Bed sediment size was an important predictor of macroinvertebrate community dynamics as shown in GEE analysis. The smallest sediment sizes (5-10 mm) were recorded in non-glacier-fed rivers and were most likely attributable to the less variable flow regimes associated with these river systems (Parker and Huryn, 2011; Blaen *et al.*, *in press*). Finer bed sediments may promote macroinvertebrate community development by attenuating variability in water temperature and flow velocity (Crossman *et al.*, 2011), thereby providing refugia from habitat disturbance in hyporheic zones. Ordination analysis showed community composition was also related to discharge, which covaried with meltwater, where most species were associated with higher discharge. This would initially appear counter-intuitive given that non-glacier-fed rivers with lower flow volumes supported the greatest taxonomic richness. However, longitudinal increases in discharge, abundance and diversity between site V1 adjacent to the glacier snout and sites V2 and BR further downstream may have had a disproportionate effect on the ordination analysis which masked finer community variations in non-glacierized rivers. Similarly, high algal growth at site V1 contrasted with lower growth further downstream and may explain the negative relationship found by GEE analysis between alpha diversity and chlorophyll *a* concentration. However, this relationship could also be a product of top-down control by chironomid communities regulating rates of algal accrual in warmer rivers (*cf.* Sturt *et al.*, 2011), although such effects have not been documented previously in Arctic rivers.

Strong relationships between water temperature and alpha and Shannon diversity demonstrated the importance of water temperature in structuring riverine communities in this area of Svalbard. Such relationships suggest water temperature plays a more significant role in determining community structure than water source contribution, although both variables are interlinked to some extent (Blaen *et al.*, *in press*). While variations in water source reflect influences on river habitat conditions (Brown *et al.*, 2003), water temperature is a more physiologically-meaningful variable which has been considered previously to be one of the most important environmental variables in explaining patterns in the distribution and abundance of chironomid taxa in European glacier-fed headwater rivers (Castella *et al.*, 2001; Lods-Crozet *et al.*, 2001; Milner *et al.*, 2001; Friberg *et al.*, 2013). The results of this study imply that this may also be the case for both glacial and non-glacial rivers in Svalbard, as inferred previously by Lods-Crozet *et al.* (2007).

Macroinvertebrate abundance at glacier-fed sites (409-993 individuals m^{-2}) was similar to those observed in other Arctic headwater rivers. Abundances at non-glacier-fed sites (259-1193 individuals m^{-2}) were substantially lower, but in line with a previous study in Svalbard (Lods-Crozet *et al.*, 2007). For example, Parker and Huryn (2011) reported mean densities of 356 ± 126 individuals m^{-2} for a glacier-fed river in Alaska, but up to 67781 ± 16740 individuals m^{-2} for a spring-fed river. Similarly, Friberg *et al.* (2001) reported densities of <20 individuals m^{-2} in a glacier-fed river in Greenland but over 4000 individuals m^{-2} for a groundwater-fed stream. Therefore, low macroinvertebrate abundance and diversity appears to be ubiquitous in Arctic rivers with high glacial inputs because few species are adapted to the cold water temperature and high disturbance levels which characterise these environments (Milner *et al.*, 2001). However, whereas Milner *et al.* (2001) predicted successive colonisation of taxa with increasing water temperature, taxa in this study were solely Chironomidae, Oligochaeta and Collembola even in relatively warm non-glacier-fed rivers (i.e. >5ºC). The limited abundance and diversity observed in non-glacier-fed rivers in Svalbard most likely reflects the harsh winter conditions and biogeographical isolation of this high-latitude archipelago (Coulson *et al.*, 2002; Prowse *et al.*, 2006). Large annual air and water temperature fluctuations and a short meltwater season impose strong selection forces on Arctic riverine biota (Coulson, 2000; Brittain *et al.*, 2009). Moreover, unlike some Alaskan Arctic spring-fed systems (Parker and Huryn, 2011), surface runoff on the Brøgger Peninsula was absent during the monitoring period from late September to early June (P. Blaen, unpublished data). Seasonal break-up and freeze-up events are highly dynamic and exert strong physical controls on river systems (Prowse and Brown, 2010). Accordingly, the diversity of many Arctic benthic communities is relatively low because few species possess the necessary traits required to survive in such conditions (Huryn *et al.*, 2005; van der Wal and Hessen, 2009).

7.4.3 Water source and water temperature optima for benthic macroinvertebrates

Meltwater contribution optima for many taxa were defined clearly, leading to the acceptance of the third study hypothesis. Over half the taxa collected had meltwater contribution optima >50%, reflecting the strong influence of snow and ice runoff on river ecosystems in Svalbard. Taxa preferences for water temperature were also clear despite the relatively low range of water temperature observed in the study (mean: 0.7-7.2 ºC). Several Diamesinae taxa, notably *D. cinerella* gr./*zernyi* gr., exhibited preferences for high meltwater and colder habitat conditions associated with sites in glacierized basins, supporting other studies which have documented this sub-family as among the first colonisers of rivers following glacial recession (Malard *et al.*, 2003; Lencioni, 2004; Brown and Milner, 2012). In contrast, *Paratanytarsus* spp. and some Orthocladiinae taxa were found only in the warmer rivers. A greater range of thermal tolerance was observed among species with lower temperature optima, suggesting that species with lower optima were tolerant of a wider temperature range than those found in warmer rivers. Consequently, glacier-fed rivers in this region of Svalbard are unlikely to be unique habitats in harbouring cold-adapted stenothermic taxa that enhance regional biodiversity, as observed elsewhere in the world (e.g. Brown *et al.*, 2007a; Finn *et al.*, 2012), because such organisms are likely to be tolerant of thermal conditions in both glacier-fed and non-glacier-fed rivers. For example, the temperature optimum for *D. cinerella*/*zernyi* gr. (2.6 ±2.0 ºC) indicates these species are capable of surviving under the water temperature regimes found in many rivers in non-glacierized river basins. Moreover, Castella *et al.* (2001) suggested that the temperature optimum for European *Diamesa* is as high as 6-7 ºC. Given that many insect populations in high-latitude regions are believed to be living in conditions cooler than their physiological optima (Deutsch *et al.*, 2008), future climate warming and associated changes in river habitat conditions could result in greater abundance and alpha diversity among benthic macroinvertebrate communities in this region of Svalbard.

7.5 Conclusions and implications for high-latitude freshwater benthic macroinvertebrate communities under a warming climate

This study investigated river habitat conditions and benthic macroinvertebrate community structure across a water source gradient in Arctic rivers to assess how climate-induced changes in hydrology may affect these river ecosystems in future. Meltwater contributions to total river flow were related significantly to several habitat variables (Hypothesis I), suggesting that water source dynamics influence discharge, pH, channel stability and the ionic composition of river water in Svalbard. Benthic macroinvertebrate taxonomic richness and abundance were related to water temperature, pH, bed sediment size and chlorophyll *a* concentration (Hypothesis II), and multivariate ordination indicated that river discharge also played an important role in determining macroinvertebrate community structure. Rivers in non-glacierized basins typically supported more diverse and abundant macroinvertebrate communities than those sourced primarily from glacial runoff. Results suggested that macroinvertebrate taxa exhibited preferences for particular levels of water source contribution (Hypothesis III), and a similar estimation of temperature optima indicated that many taxa in glacier-fed rivers were also tolerant of the thermal conditions found in rivers in non-glacierized basins.

In the context of ongoing glacier recession and future warming in Arctic regions (Anisimov *et al.*, 2007; Zemp *et al.*, 2009), declining meltwater inputs and increased permafrost wasting are expected to result in a progressive shift towards groundwater-dominated river systems (Smith *et al.*, 2007; Walvoord and Streigl, 2007; St. Jacques and Sauchyn, 2009). The results of this study support others in suggesting these changes in hydrology will modify flow regimes in some high-latitude riverine environments (Hood and Berner, 2009; Milner *et al.*, 2009; Blaen *et al.*, *in press*), which in turn may result in a general increase in water temperature and channel stability. Together, these are likely to lead to an increase in the abundance and diversity of macroinvertebrate communities in some Arctic rivers. However, changes in aquatic community structure in Svalbard are likely to be mediated by harsh winter conditions and biogeographical constraints on biotic interactions (Coulson *et al.*, 2000), and therefore the response of river ecosystems to climate change on this high-latitude archipelago may be different to those in other Arctic regions.

7.6 Chapter summary

Linkages between dynamic water source contributions, habitat variables and macroinvertebrate communities in Svalbard river basins have been assessed. The results highlight the strong influence of hydrology and habitat variables on aquatic biodiversity in these high-latitude rivers and provide a process basis on which to model changes in Arctic riverine biodiversity under future environmental conditions. The findings of Chapters 4-7 are synthesised in Chapter 8 to provide a more holistic understanding of the response of Arctic river systems to climate change.

CHAPTER 8

SYNTHESIS AND FUTURE DIRECTIONS

8.1 Introduction

The overarching aim of this study was to adopt an interdisciplinary approach to understand process connections between water source partitioning, aquatic habitat conditions and ecological structure and functioning in Arctic river basins, and thus assess the vulnerability of these systems to climate change. An overview of ongoing and future predicted environmental change in the Arctic was presented in Chapters 1 and 2 which emphasised the sensitivity of cryospheric and hydroecological linkages in Arctic river systems to climate variability. Key knowledge gaps in our contemporary understanding of Arctic river systems were identified for further research which focused on water source (Chapter 4), water temperature (Chapter 5), nutrient (Chapter 6) and ecological (Chapter 7) dynamics. In this concluding chapter the major findings of the research undertaken are summarised holistically and a synthesis in the context of future environmental change in Arctic regions is provided. Finally, potential areas for further research on high-latitude northern river ecosystems are suggested and key contributions of the research to Arctic hydroecological understanding are presented.

8.2 Major research findings

The major outcomes of the research were as follows:

1. Hydrochemical and isotopic characteristics of rivers in Svalbard reflected spatially-distinct hydrological processes operating over diurnal to seasonal timescales. Their detailed study enabled the identification of key basin water sources. Quantification of seasonal and year-to-year variations in water source contributions to bulk river flow demonstrated the importance of glacial runoff in sustaining river discharge following snowmelt, but also indicated that seasonal thawing of the active layer led to increased inputs from shallow nutrient-enriched groundwater systems in late summer.

2. Water temperature in Svalbard rivers was driven by a combination of water source partitioning, exposure time to the atmosphere, prevailing meteorological conditions, river discharge, runoff interaction with permafrost and buried ice, and basin-specific geomorphological features. The findings suggest warmer and more variable thermal regimes may prevail in high-latitude rivers in future as the proportion of bulk discharge sourced from glacial meltwater declines and rivers shift towards greater snow- and groundwater contributions to flow.

3. Demand for NH_4^+ and acetate was high in headwater rivers in the Kongsfjord region of Svalbard. Uptake of NH_4^+ was associated positively with water temperature, algal biomass and transient storage area, while acetate uptake was associated positively with sediment respiration and more stable river channels. Uptake of NH_4^+ increased when added with acetate, suggesting NH_4^+ assimilation by microbial communities is mediated by labile DOC availability. Consequently, future changes in the nutrient export from these river basins may be controlled by both physical and biological process interactions.

4. Meltwater contributions to total river flow were related significantly to several physicochemical habitat variables, suggesting future shifts in water source dynamics are likely to influence aquatic habitat conditions in Svalbard rivers. Water temperature was associated positively with benthic macroinvertebrate taxonomic richness, indicating that warmer rivers in non-glacierised basins may play an important role in supporting regional macroinvertebrate biodiversity in some Arctic areas.

8.3 Synthesis

8.3.1 Climate change in the Arctic

The high climatic sensitivity and low human perturbation of Arctic river basins make these systems vital indicators of global environmental change by providing insight into hydroecological responses to high-latitude climate forcing (Nelson *et al.*, 2002; Ye *et al.*, 2003; Milner *et al.*, 2009). There is now a broad consensus that the Arctic is warming more rapidly than the global average and that this trend, coupled with increased precipitation, is likely to continue through the remainder of the 21st Century (Kattsov *et al.*, 2005; Anisimov *et al.*, 2007). These climatic changes are expected to have a major influence on cryospheric systems (Walsh *et al.*, 2005; Comiso *et al.*, 2008; Serreze et al, 2009; Callaghan *et al.*, 2011) which will alter hydrological processes in these environments (Hinzman *et al.*, 2005). Changes in river flow regimes are likely to modify physico-chemical habitat conditions and have implications for biodiversity and ecosystem processes in Arctic rivers (White *et al.*, 2007; Hannah *et al.*, 2007; Milner *et al.*, 2009). Yet, many high-latitude river ecosystems have remained relatively understudied, with the majority of hydroecological research focused on systems in North America (e.g. Slavik *et al.*, 2004; Huryn, 2005; Parker and Huryn, 2011). Consequently, this highly interdisciplinary study has provided new process-based knowledge of river ecosystems in Svalbard and creates a foundation for assessing the hydroecological response of these Arctic systems to global environmental change.

8.3.2 Water sourcing in Arctic river basins

Contemporary understanding of Arctic basin functioning was enhanced by identifying the key processes governing Svalbard flow regimes (Chapter 4). Modelling of end-member mixing demonstrated dilute meltwater dominated river flows, particularly in glacierized basins, but that groundwater inputs increased in late summer with deepening of the active soil layer. These results inform a new conceptual diagram (Figure 8.1a) to illustrate current seasonal patterns in the relative contribution of water sources to flow in a hypothetical glacierized Arctic river basin, building on the previous version in Chapter 2. Key differences between the two diagrams are: (1) improved understanding of how glacier surface runoff sustains river flow throughout the melt season, (2) inclusion of rainfall-runoff as an additional water source, (3) recognition that supra-permafrost groundwater contributions are low and that sub-permafrost groundwater discharge is often virtually absent in regions of continuous permafrost.

Annual river discharge is likely to decrease with glacier recession (Barnett *et al.*, 2005) although this may be preceded by a peak in icemelt runoff over a short to medium timeframe for larger glaciers (Figure 8.1b; Milner *et al.*, 2009). Forecasted increases in winter snowfall are likely to enhance the magnitude of the spring freshet as large volumes of snowmelt runoff are released to river channels (White *et al.*, 2007), while the role of non-glacial water sources in maintaining river flows could become increasingly important as glaciers shrink and permafrost systems degrade (Lappegard *et al.*, 2007; Walvoord and Streigl, 2007; Adam and Lettenmaier, 2008; Moore *et al.*, 2009). Moreover, increased summer precipitation in Arctic regions (Kattsov *et al.*, 2007; Overland *et al.*, 2011) is likely to increase the frequency of short intense flood events where rainfall-runoff dominates river flow (Dugan *et al.*, 2009).

Long-term future climate change in Arctic regions may lead to the dominance of river flows by additional water sources that were not evaluated (Figure 8.1c). For example, warming at high latitudes could alter the basal thermal regime of valley glaciers (Rippin *et al.*, 2003), resulting in the development of distributed drainage structures where sub-glacial meltwaters acquire unique hydrochemical signatures distinct from both dilute supra-glacial runoff and ionically enriched groundwater (Hodson *et al.*, 2002; Tranter *et al.*, 2002; Brown *et al.*, 2006). The prevalence of sub-permafrost groundwater discharge and sub-glacial distributed drainage systems may increase in future in association with glacier thinning and reduced permafrost coverage (Sharp *et al.*, 2011; Lawrence and Slater, 2005). Finally, rain-on-snow events are forecasted to become more common in some

Arctic areas during the following century (Putkonen and Roe, 2003), and more frequent thaw events could result in the release of snowpack meltwater to rivers during winter months (Callaghan *et al.*, 2011c). These changes in runoff regime may require the development of more advanced hydrological models in future to achieve more accurate predictions of river flow regimes (e.g. Brown *et al.*, 2006; Cable *et al.*, 2011).

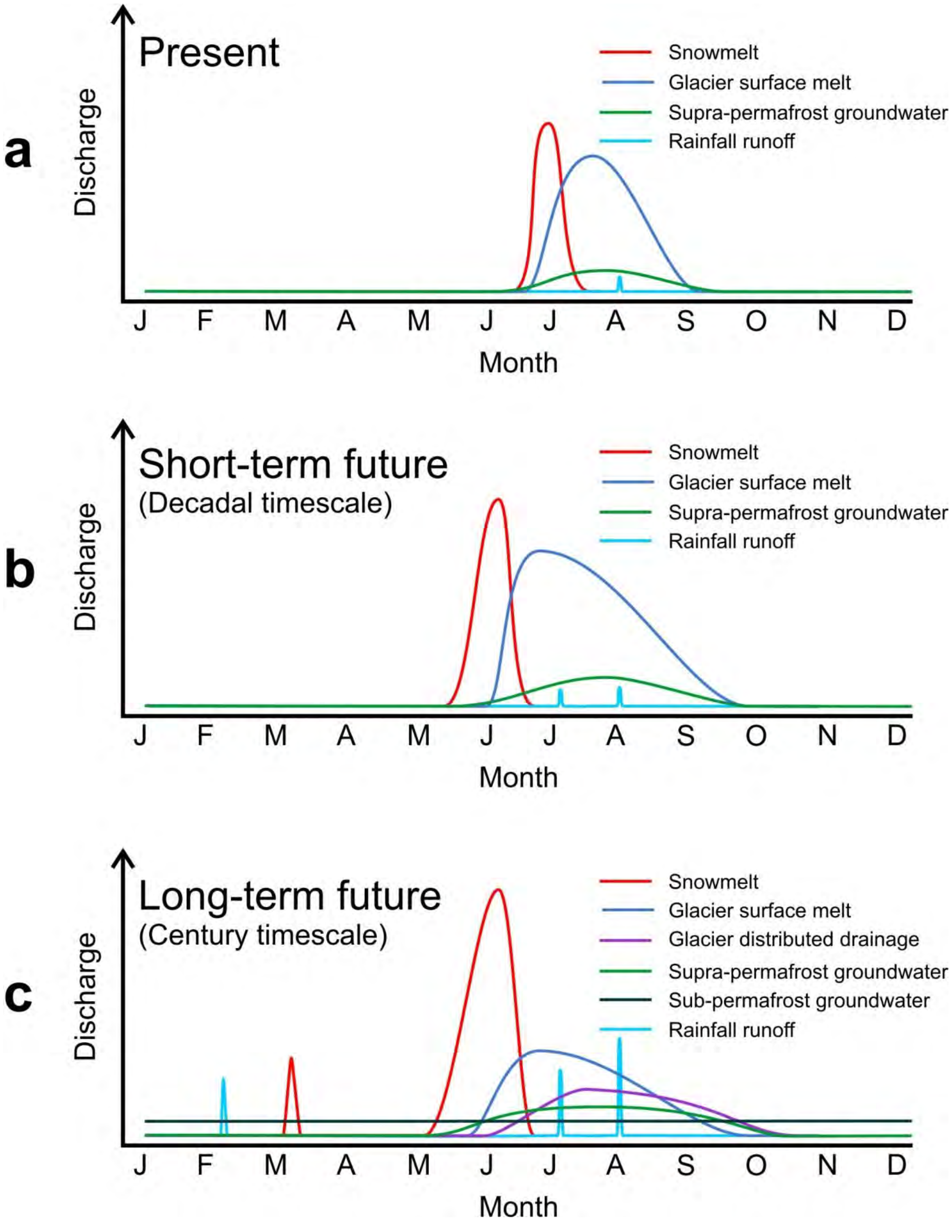


Figure 8.1: *Hypothesised present and future timing and magnitude in relative levels of water source contribution to river discharge in an Arctic glacierized basin*

8.3.3 Water temperature dynamics in Arctic rivers

As outlined in Chapter 4, water source dynamics exert strong control on river flow regimes, indicating future changes in relative contributions of meltwater and groundwater might affect aquatic habitat conditions in Arctic rivers. This concept was explored in greater detail in Chapter 5 by focusing on one habitat variable, water temperature, which has been shown previously to be a strong influence on glacierised headwater community structure and function (Milner *et al.*, 2001; Brown *et al.*, 2007b). Water temperature in rivers sourced primarily from glacier meltwater was colder and less variable than river water temperature in non-glacier-fed rivers, demonstrating water source to be a substantial driver of these river thermal regimes. Additional drivers of water temperature in these high-latitude basins included incoming solar radiation, basin geomorphology and, in the case of glacier-fed rivers, distance from source.

River thermal regimes cannot be predicted accurately from water source alone. Nevertheless, this research offered some insights to changes in water temperature in Arctic rivers under a changing climate. Warmer and more variable water temperature regimes may be expected in future as surface air temperature increases and rivers become increasingly dominated by non-glacial water sources (Anisimov *et al.*, 2007; Walsh *et al.*, 2011; Milner *et al.*, 2009). Glacier retreat is likely to limit the coldest river temperature conditions to headwater areas, while thermokarsting associated with permafrost degradation could also create new bodies of standing water (Woo *et al.*, 2008) with relatively high outflow temperatures (*cf.* Hieber *et al.*, 2002). However, these potential rises in river temperature might be mitigated by changes in cloud cover in high-latitude regions which modify incoming short-wave radiation (Oke, 1987; Kattsov *et al.*, 2005), while in the near future warmer groundwater inputs may be offset by an increase in glacier meltwater production in response to climate forcing (Milner *et al.*, 2009).

Finally, the deepening of active soil layers in proglacial areas could reduce interaction of groundwater with buried ice (Engeset and Weydahl, 1998; Cooper *et al.*, 2002). Permafrost is believed to play a major role in mediating groundwater temperature in Svalbard because bed temperature cooled with depth at all monitored sites (e.g. Figure 8.2), signifying downwelling of surface waters through the hyporheic zone (*cf.* upwelling of warmer groundwater; Brown *et al.*, 2006). These results have implications for river benthic macroinvertebrate communities because hyporheic zones are often considered to provide thermal refugia to biota, as observed previously in alpine glacierised river basins (Malard *et al.*, 2001; Cadbury *et al.*, 2008). Although bed temperature in Svalbard was lower generally than river water temperature during the monitoring period, thermal dynamics in river beds exhibited a damping of response signal with depth, suggesting hyporheic zones may mitigate cold surface temperature conditions and facilitate the persistence of benthic macroinvertebrate communities. However, it is unknown whether water flow through river bed sediments exists on a perennial basis in Svalbard, as in other Arctic regions (e.g. Prowse *et al.*, 2006; Parker and Huryn, 2011) or whether these subsurface habitats are frozen entirely during winter months.

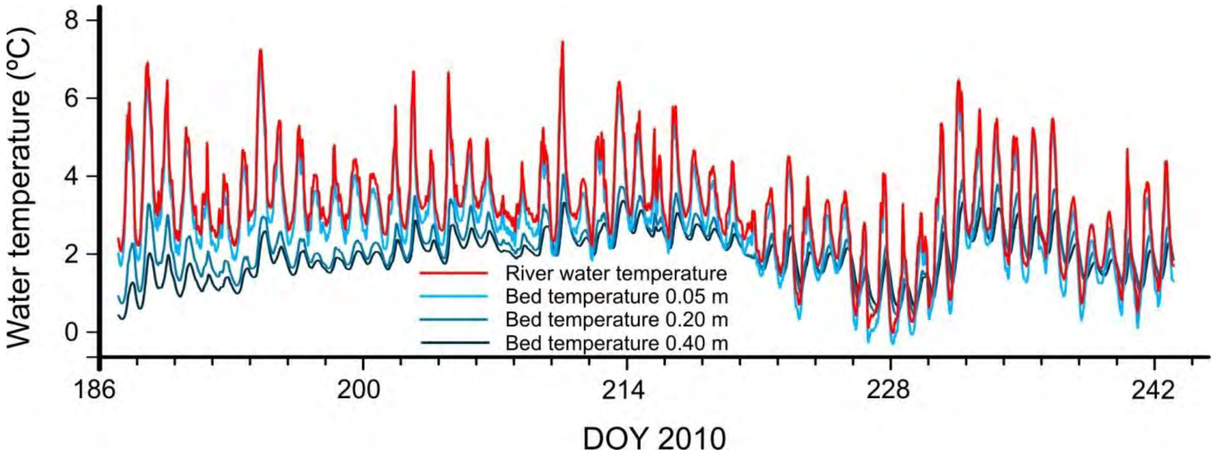


Figure 8.2: *Water and bed temperature dynamics at site V2 during the 2010 melt season*

8.3.4 Nutrient cycling in Arctic river basins

A basis for analogue modelling of the impact of climate-induced changes in water temperature in Svalbard rivers during the 21st century was provided in Chapter 5. Water temperature was found to be a key variable in controlling rates of ecosystem functioning and aquatic community structure (Chapters 6 and 7). The processes by which future changes in environmental habitat conditions and the supply of nutrients to Arctic rivers might affect rates of nutrient uptake in these environments were explored in Chapter 6. A comparison of NH_4^+ uptake velocities in the study rivers with previously published values (Figure 8.3) demonstrates that despite biologically unfavourable environmental conditions (i.e. cold water temperature, strong seasonality, low channel stability), rates of biogeochemical processing in some Svalbard rivers are comparable to those recorded in rivers at lower-latitudes. Unfortunately, it is not possible to achieve a similar comparison of global acetate uptake velocities due to a lack of comparable studies.

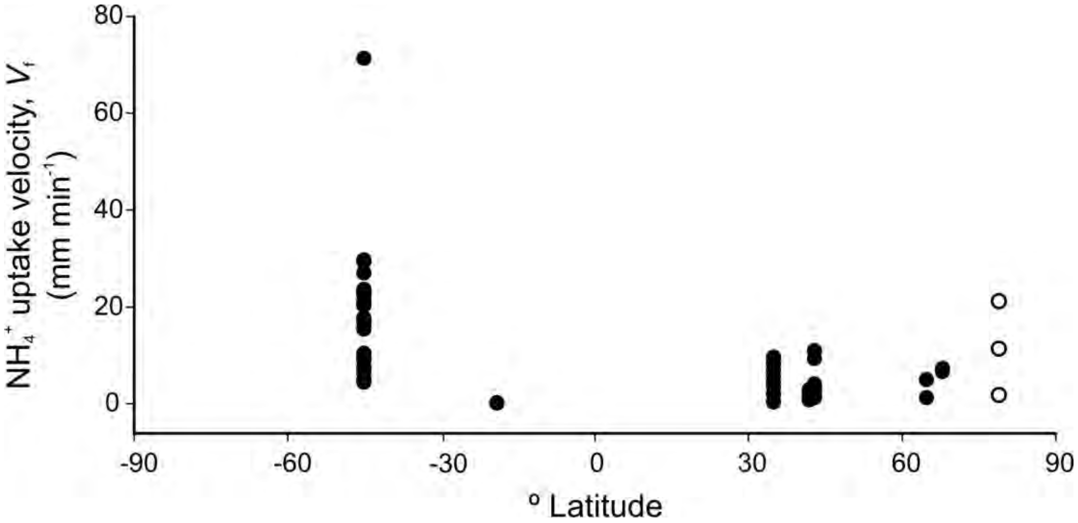


Figure 8.3: *Comparison of global NH_4^+ uptake velocities (data from Ensign and Doyle (2006), Rasmussen et al. (2011) and the present study (open circles))*

Demand for NH_4^+ was associated positively with water temperature, suggesting future trends towards warmer thermal regimes may enhance the capacity for nutrient uptake in these rivers (Demars *et al.*, 2011). Warmer conditions will increase the depth of seasonal active soil layers, increasing groundwater discharge and mobilising potentially large pools of nitrogen and carbon stored in high-latitude tundra ecosystems (Frey and Smith, 2005; Petrone *et al.*, 2006). Groundwater seeps at the beginning of melt seasons (see Chapter 6) were not significantly enriched in most nutrients (except PO_4^-) relative to meltwater. However, seasonal monitoring of water source hydrochemistry (Chapter 4) showed groundwater samples contained relatively high concentrations of NH_4^+, NO_3^-, and most other major ions, suggesting long-term leaching of water from proglacial soils has the potential to enhance river nutrient concentrations (Gersper *et al.*, 1980; Hodson *et al.*, 2002). In addition, isotopic data (Chapter 4) indicated episodic rainfall events in Svalbard river basins can displace water from shallow groundwater systems. Increased future precipitation forecasted for Arctic regions (Kattsov *et al.*, 2006) may enhance flushing of water from soil layers and thereby enhance nutrient loading in river channels. Therefore, while water source contributions were not linked directly to nutrient uptake values, this research has provided a more mechanistic understanding of nutrient cycling in Arctic rivers by identifying potential drivers of nutrient uptake which might be modified by basin-scale hydrological changes associated with climate forcing.

Further influences on NH_4^+ demand were identified as algal biomass, transient storage area and DOC availability. In future, warmer water temperature and increased nutrient availability could promote algal growth in Arctic rivers, leading to an intensification of nutrient cycling in these environments. However, despite concerns that continued warming could affect thaw bulb depths (Edwardson *et al.*, 2003; Merck *et al.*, 2012), such changes appear unlikely to affect hyporheic transient storage properties because estimated values (Chapter 6) were similar to those observed previously in low-latitude temperate rivers (e.g. Scordo and Moore, 2009). The finding that DOC availability may moderate high-latitude N cycling supports previous research which has suggested the majority of DOC in northern rivers is recalcitrant for much of each year (Dittmar and Kattner, 2003; Holmes *et al.*, 2008), although permafrost degradation and increased snowmelt volumes may result in a higher fraction of labile DOC in some Arctic rivers (O'Donnell *et al.*, 2010). Consequently, future rates of high-latitude freshwater nutrient cycling are likely to be controlled by both physical and biological process interactions which could be modified by changing hydrological conditions.

8.3.5 Hydrological drivers of macroinvertebrate communities in Arctic rivers

Research in alpine basins has shown rivers comprised primarily of groundwater to be associated with less harsh habitat conditions than glacially-dominated systems (Brown *et al.*, 2003; Brown *et al.* 2006). These conditions are reflected typically by higher taxonomic diversity and abundance (Ilg and Castella, 2006; Brown *et al.*, 2007b). Water source-habitat-biota relationships were less clearly defined in this study: some habitat variables (e.g. EC, channel stability) were related significantly to water source but these were not always the best predictors of macroinvertebrate community structure. This discrepancy may arise from additional environmental drivers which override the influence of water source on habitat conditions in some Arctic rivers. For example, rapid longitudinal warming of river temperature away from glacier snouts (Chapter 5) was due to a combination of both water source and atmospheric energy receipt in proglacial areas. However, despite these complexities, observed differences in physicochemical habitat variables between rivers in glacierised and non-glacierised basins provided insight into potential future conditions that might be expected under scenarios of future climate change and continued glacier retreat (Anisimov *et al.*, 2007; Sharp *et al.*, 2011; Walsh *et al.*, 2011). Rivers in non-glacierised basins were characterised by lower specific discharge, SSC and flow variability but higher water temperature, EC, nutrient concentration and channel stability than in glacier-fed rivers. The results of this study suggest that a progressive shift towards rivers dominated by non-glacial water sources in future (Walvoord and Streigl, 2007; Adam and Lettenmaier, 2008) may increase the availability of warmer, less variable and more nutrient-rich habitat conditions in Arctic rivers.

These projected changes in physicochemical habitat characteristics during summer months are envisaged as a function of season and water source (Figure 8.4) and related to the harsh-benign concept of Peckarsky (1983). At present, habitat conditions in Svalbard rivers provide one short annual _window of ecological opportunity‘ for riverine biota each year which occurs during the summer flow period. Freezing conditions during winter months create harsh habitat conditions which limit growth and development of individuals during this time (Huryn *et al.*, 2005). In future, warming in Arctic regions (Anisimov *et al.*, 2007; Overland *et al.*, 2011; Vavrus *et al.*, 2012) may lengthen the duration of seasonal river flow periods (Prowse and Brown, 2010). Increased snow accumulation in winter months (Callaghan *et al.*, 2011) could increase river habitat disturbance where snowmelt runoff dominates spring flows (White *et al.*, 2007). In addition, the apparent harsh-benign habitat gradient between meltwater- and groundwater-fed rivers observed at present may be complicated in future as summer rainfall events create short pulses of disturbance in these high-latitude ecosystems (Jones and Rinehart, 2010; Milner *et al.*, 2012).

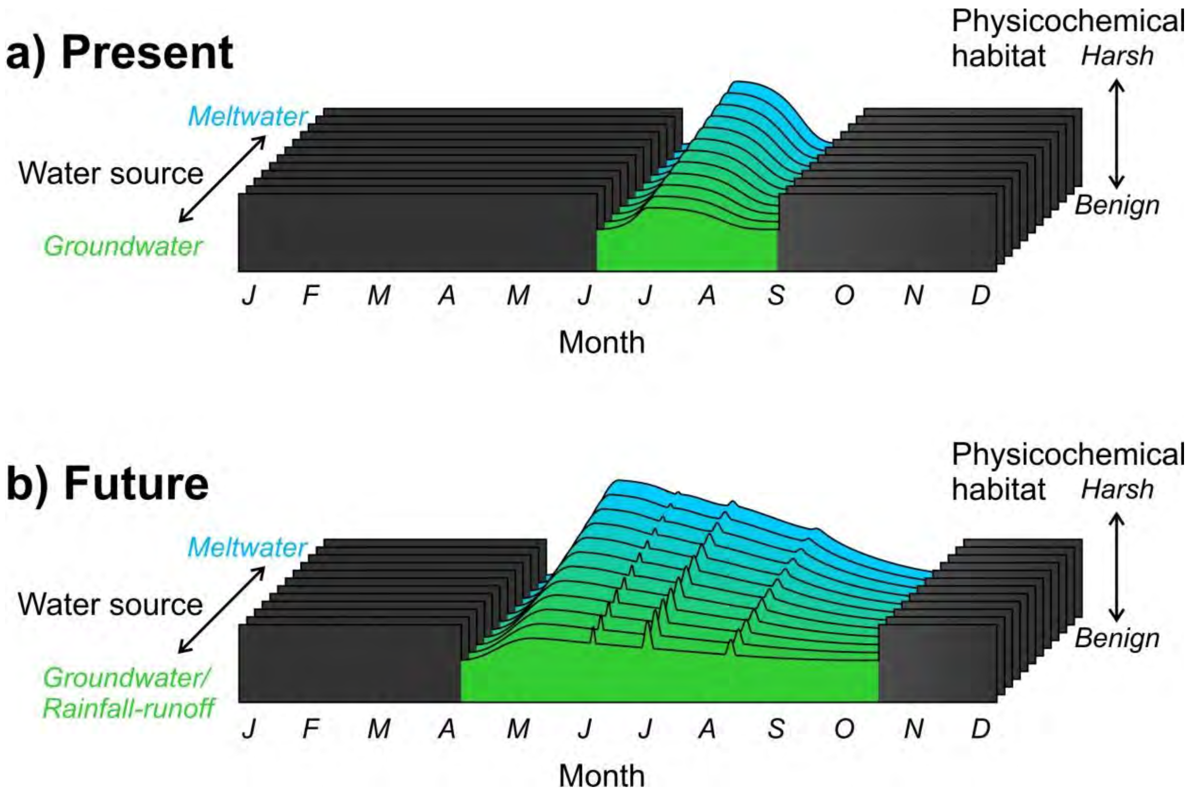


Figure 8.4: *Conceptual model outlining seasonal changes in physicochemical habitat condition along a water source gradient in Arctic rivers for a) present conditions and b) future conditions. Dark shading indicates periods where surface river flow is absent.*

Benthic macroinvertebrate taxa richness was highest in non-glacier-fed rivers and some species only found in these environments. Riverine macroinvertebrate communities in Arctic and alpine basins are structured by a combination of habitat variables, but research in Chapter 7 showed water temperature to be particularly important for determining benthic macroinvertebrate taxa richness in these Arctic rivers (Milner *et al.*, 2001). Consequently, forecasted warming at high latitudes could lead to increased biotic diversity in these river systems if regional taxa pools exceed local diversity (Milner *et al.*, 2001; Brown *et al.*, 2007a; Parker and Huryn, 2011). However, the composition of future benthic macroinvertebrate communities is unlikely to be entirely predictable because community assembly processes are believed to become more stochastic at higher temperature (Milner *et al.*, 2008; Brown and Milner, 2012). Increased warming could also increase species abundance given that macroinvertebrate developmental cycles are often temperature-regulated (Danks, 2007) and many Arctic species are believed to be living in conditions cooler than their physiological optima (Deutsch *et al.*, 2008).

In contrast to alpine river research, previous studies concerning the impact of warming on Arctic aquatic communities have been equivocal in their conclusions. Many have emphasised the physiological importance of water temperature in driving high-latitude river biodiversity and ecosystem functioning (Friberg *et al.*, 2009; Demars *et al.*, 2011; Friberg *et al.*, 2013). However, climate change models generally forecast the highest temperature rises in winter (Overland *et al.*, 2011), and therefore some studies have suggested that future changes in aquatic community composition will be controlled to a greater extent by changes in habitat disturbance related to permafrost degradation and shifts in spring ice breakup intensity (Durand *et al.*, 2011). This study supports broadly the findings of Milner *et al.*‘s (2001) pan-European study of glacierised headwater rivers which indicated water temperature and channel stability (a proxy measure of habitat disturbance) to be important drivers of biotic community composition in high-latitude rivers. These habitat variables are highly likely to be modified in future by hydrological variability arising from climate and cryospheric changes. Given the freezing winter conditions that characterise much of the Arctic (McBean *et al.*, 2005), it is improbable that even substantial warming will generate perennial flows in most areas and therefore northern regional taxa pools are likely to remain restricted to species with traits for freeze tolerance or avoidance for the foreseeable future (Huryn *et al.*, 2005; van der Wal and Hessen, 2009). Moreover, biogeographical dispersal constraints impose further limitations on benthic macroinvertebrate community composition within the Svalbard archipelago (Coulson *et al.*, 2002). The contrast in benthic macroinvertebrate diversity between continental European glacier-fed rivers and those in Svalbard is highlighted graphically (Figure 8.5). The conceptual model demonstrates the low biotic diversity and abundance observed among macroinvertebrate communities in this study and shows how additional taxonomic families do not colonise even relatively warm and stable river habitats in north-west Svalbard due to biogeographical constraints.

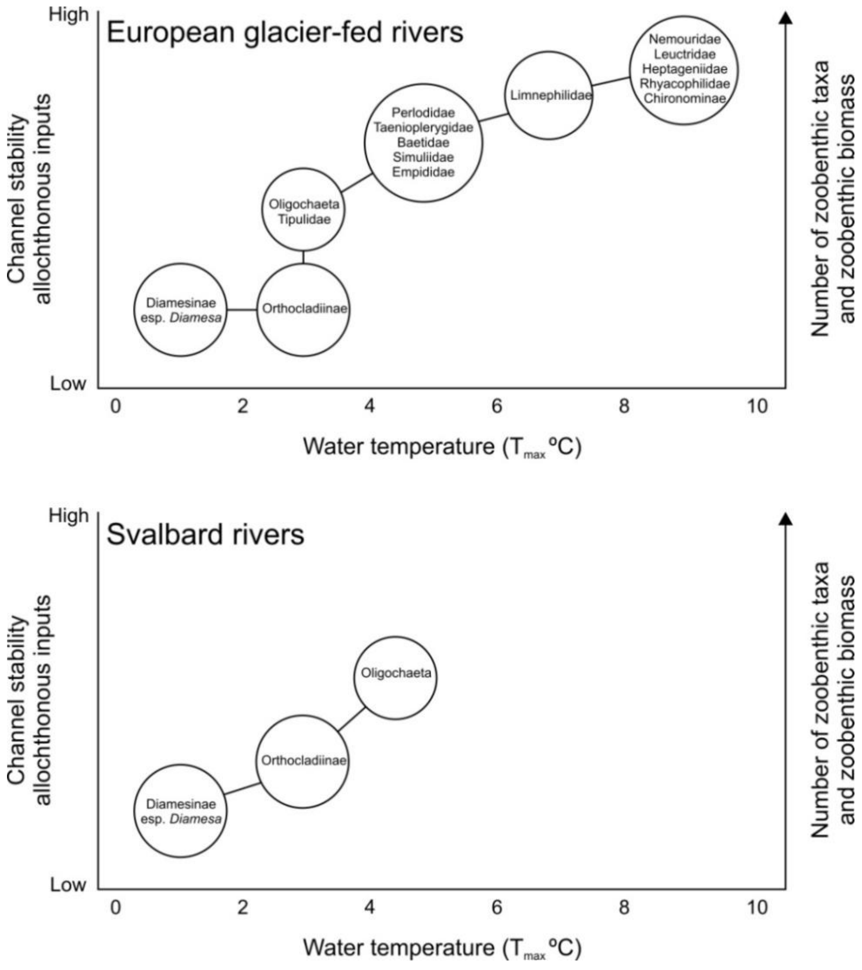


Figure 8.5: *Conceptual model describing the likely first appearance of benthic macroinvertebrate taxa in European glacier-fed rivers and rivers in the Kongsfjord region of Svalbard as a function of water temperature and channel stability.*
(modified from Milner et al., 2001)

8.3.6 Summary

An integrated approach was taken to identify linkages between water sourcing, aquatic habitat conditions and ecological structure and functioning in river basins in Svalbard. These systems have until now been relatively understudied, despite their evident vulnerability to climate forcing and the potential for rapid environmental change in northern regions (McBean *et al.*, 2005; Overland *et al.*, 2011). Consequently, this research has improved contemporary knowledge of Arctic river ecosystems by providing a more holistic understanding of their hydroecology. A model of hypothesised links between atmospheric and cryospheric processes and Arctic river hydroecology was first introduced in section 2.5. The model has been refined based on the research outcomes of this study (Figure 8.6). Key differences between the new and original models are: (1) the role of basin water stores in controlling hydrological and nutrient fluxes; (2) improved understanding of the linkages between water sourcing and environmental habitat variables; (3) incorporation of additional influences on Arctic hydroecology (e.g. basin geology and local geomorphology). The model provides a graphical representation of the linkages between environmental system components in Arctic river basins and thus provides a framework on which to assess the response of Arctic river ecosystems to climate change.

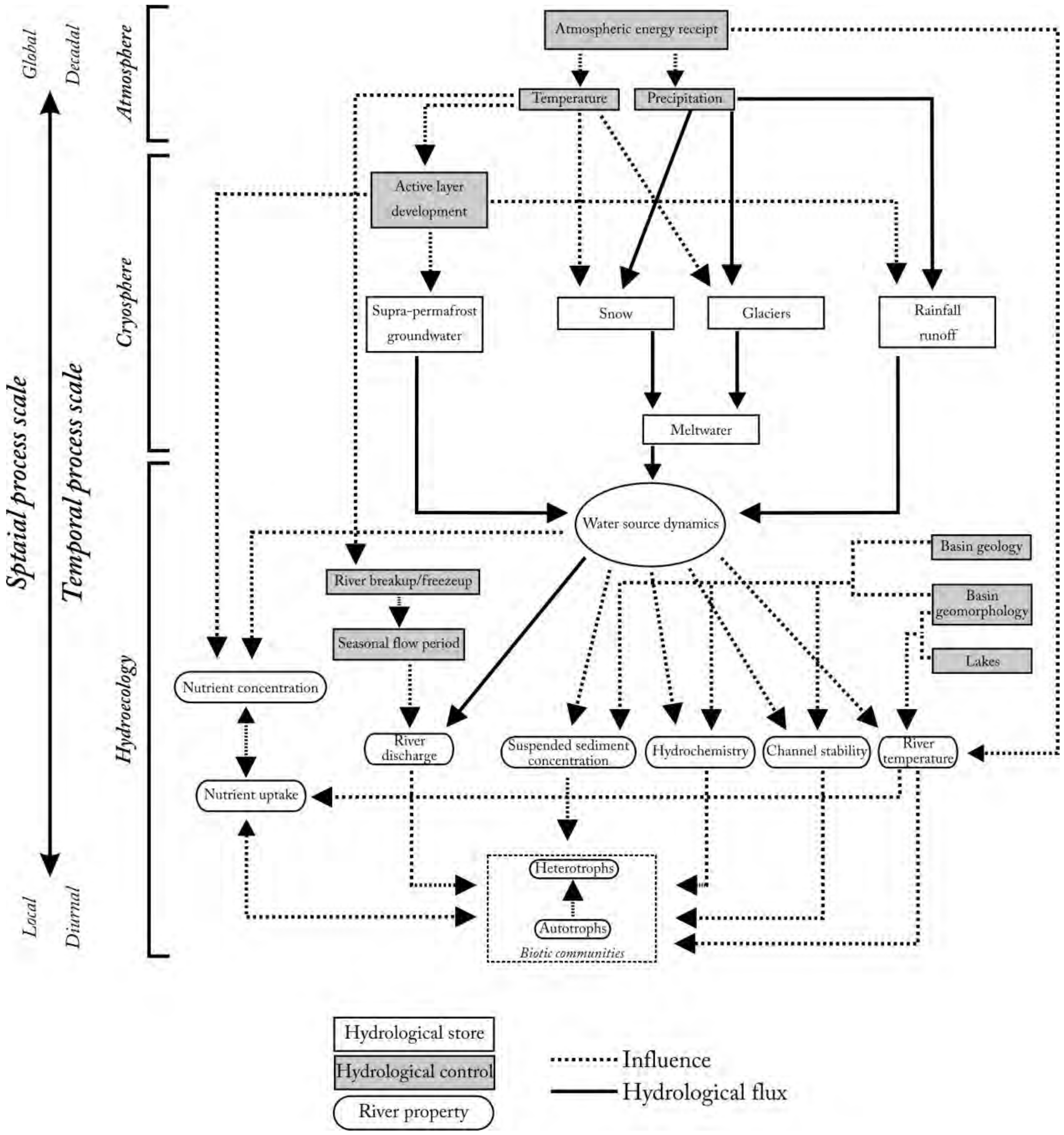


Figure 8.6: *Conceptual model of climate-cryosphere-hydroecology links in Arctic river basins operating at different scales. Arrows demonstrate theoretical relationships between components. (Revised from Chapter 2)*

8.4 Recommendations for further research

Areas where further research is now warranted on the basis of the research findings of the study are outlined in this section. These ideas may be useful in extending contemporary understanding of the drivers of hydroecological processes in Arctic river basins.

8.4.1 Water sourcing in Arctic river basins

Hydrochemical sampling to inform end-member mixing modelling was conducted on a weekly basis (Chapter 4). This sampling resolution was appropriate to assess seasonal trends in water source dynamics. However, a more frequent sampling framework would provide understanding of sub-daily variations in water source contributions to river flow driven by diurnal cycles in atmospheric energy receipt. High-resolution sampling could also promote greater awareness of episodic rainfall events and their influence on hydrological processes and flow pathways (e.g. Brown and Hannah, 2007). Additionally, further insight to the influence of permafrost on basin-scale hydrological behaviour might be gleaned from the installation of a piezometer network to better understand groundwater-surface water interactions and subsurface flow pathways (Brosten *et al.*, 2009; Dragon and Marciniak, 2010). Finally, despite the inherent challenges of working at high latitudes (Hodgkins *et al.*, 2009), there remains a need to characterise the hydrology of break-up and freeze-up processes in Arctic rivers given that the timing and duration of the active flow season is likely to be altered under future climate change scenarios.

8.4.2 Water temperature dynamics in Arctic rivers

Patterns in Svalbard river water temperature and links to controlling variables were explored in Chapter 5. As with basin water sourcing, the integration of these data with piezometer measurement would increase contemporary understanding of how groundwater temperature is modified by interaction with permafrost layers. Further research into river bed temperature dynamics, particularly during winter months, could provide insight into overwintering habitat conditions for aquatic biota. Additionally, the coupling of water temperature data with additional atmospheric information from local automatic weather stations would allow for the quantification of energy fluxes that drive river heat budgets (e.g. Cozzetto *et al.*, 2006) and facilitate predictions of future thermal variability in these high-latitude systems. Consequently, such information might be used to determine future shifts in freshwater ecosystem structure and functioning given the strong influence of water temperature in Arctic rivers (Chapters 6 and 7).

8.4.3 Nutrient cycling in Arctic river basins

Demand for NH_4^+ in Svalbard rivers may be controlled to some extent by the availability of labile DOC in the water column (Chapter 6). Previous studies have observed large seasonal variations in the bioavailability of DOC in Arctic rivers (Holmes *et al.*, 2008). Therefore, future work in this area could focus on assessing whether nutrient demand varies temporally throughout the melt season. The methodology employed by this study was suitable for characterising nutrient uptake in several rivers within a limited budget but was not capable of discerning the specific processes that led to the removal of nutrients from the water column. This could be overcome through the use of more sensitive techniques using isotopic tracers such as ^{14}C or ^{15}N to provide detailed information on nutrient uptake mechanisms (e.g. Mulholland *et al.*, 2008), although the cost of applying these methods is considerable. Additionally, a new technique for rapid quantification of nutrient uptake parameters has been developed recently (Covino *et al.*, 2010). This may hold promise for Arctic rivers where highly variable flow regimes can cause difficulties in achieving the plateau conditions required by traditional nutrient uptake methods (Tank *et al.*, 2006).

8.4.4 Hydrological drivers of macroinvertebrate communities in Arctic rivers

Benthic macroinvertebrate community sampling (Chapter 7) was conducted at the reach scale because this aligned with similar-scale variations in water sourcing and physicochemical habitat changes. Sampling at a finer spatiotemporal scale may allow for the identification of patch-scale influences on community dynamics (e.g. interstitial space availability, flow velocity). Conversely, larger-scale analyses may be possible if ecological data from different high-latitude locations could be combined to form a pan-Arctic dataset (Overland *et al.*, 2004; Post *et al.*, 2009). Finally, as with most ecological research in Svalbard, this study focused on sampling sites in the north-west of the archipelago. Future hydroecological research might be directed towards the relatively understudied east coast where harsher climatic conditions may result in very different aquatic communities (Coulson, 2007).

8.5 Key contributions to Arctic hydroecological understanding

The main aim of this study was to conduct an interdisciplinary study to improve contemporary understanding of the hydroecology of Arctic river basins. This aim has been met by:

1. increasing contemporary understanding of spatiotemporal water source dynamics within Svalbard river basins;
2. identifying process connections between basin hydrology, physicochemical habitat conditions, ecosystem structure and ecosystem functioning within these river basins;
3. using this knowledge as a basis for assessing potential changes in Arctic river ecosystems under future environmental conditions.

This interdisciplinary hydroecological approach provides a framework to understand the vulnerability of Arctic river ecosystems to future climate change (Kattsov *et al.*, 2005; Anisimov *et al.*, 2007). However, further holistic understanding of climate-hydrology-ecology linkages in high-latitude regions is now required. Previous research has explored atmospheric influences on hydrological processes in high-latitude regions (Peterson *et al.*, 2002; McClelland *et al.*, 2004), yet no studies have elucidated relationships between atmospheric circulation patterns, high-latitude hydrology, river physicochemical habitat conditions, and aquatic ecosystem structure and functioning. The identification of such relationships would enable accurate predictions of Arctic river hydroecology to be generated from large-scale climate data.

List of references

Acuña V, Wolf A, Uehlinger U, Tockner K. 2008. Temperature dependence of stream benthic respiration in an Alpine river network under global warming *Freshwater Biology* **53**: 2076-2088.

Adam JC, Lettenmaier DP. 2008. Application of New Precipitation and Reconstructed Streamflow Products to Streamflow Trend Attribution in Northern Eurasia *Journal of Climate* **21**: 1807-1828.

Aldridge KT, Brookes JD, Ganf GG. 2010. Changes in abiotic and biotic phosphorus uptake across a gradient of stream condition *River Research and Applications* **26**: 636-649.

Alexander RB, Boyer EW, Smith RA, Schwarz GE, Moore RB. 2007. The role of headwater streams in downstream water quality *Journal of the American Water Resources Association* **43**: 41-59.

Alexander RB, Smith RA, Schwarz GE. 2000. Effect of stream channel size on the delivery of nitrogen to the Gulf of Mexico *Nature* **403**: 758-761.

Allen E, Spence D. 1981. The differential ability of aquatic plants to utilize the inorganic carbon supply in fresh waters *New Phytologist* **87**: 269-283.

American Public Health Association (APHA). 2002. *Standard Methods for the Examination of Water and Wastewater*. American Public Health Association: Washington, DC.

Amon RMW, Meon B. 2004. The biogeochemistry of dissolved organic matter and nutrients in two large Arctic estuaries and potential implications for our understanding of the Arctic Ocean system *Marine Chemistry* **92**: 311-330.

Anderson SP. 2005. Glaciers show direct linkage between erosion rate and chemical weathering fluxes *Geomorphology* **67**: 147-157.

Anderson SP, Drever JI, Frost CD, Holden P. 2000. Chemical weathering in the foreland of a retreating glacier *Geochimica et Cosmochimica Acta* **64**: 1173-1189.

Anisimov OA, Vaughan DG, Callaghan T, Furgal C, Marchant H, Prowse TD, Vilhjálmsson H, Walsh JE. 2007. Polar regions (Arctic and Antarctic) in *Climate Change 2007: impacts, adaptation and vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the*

Intergovernmental Panel on Climate Change, Parry ML, Canziani OF, Palutikof JP, van der Linden PJ, Hanson CE (eds). Cambridge University Press: Cambridge.

Armitage PD, Cranston P, Pinder LCV. 1995. *The Chironomidae: biology and ecology of non-biting midges*. Springer: UK.

Arscott DB, Tockner K, Ward JV. 2001. Thermal heterogeneity along a braided floodplain river (Tagliamento River, northeastern Italy) *Canadian Journal of Fisheries and Aquatic Sciences* **58**: 2359-2373.

ASTER Global DEM. 2010. *NASA Land Processes Distributed Active Archive Center* Available at: https://lpdaac.usgs.gov/get_data Accessed: 01/06/2012

Ávila-Jiménez ML, Coulson SJ, Solhøy T, Sjöblom A. 2010. Overwintering of terrestrial Arctic arthropods: the fauna of Svalbard now and in the future *Polar Research* **29**: 127-137.

Balestrini R, Arese C, Freppaz M, Buffagni A. 2012. Catchment features controlling nitrogen dynamics in running waters above the tree line (Central Italian Alps) *Hydrology and Earth System Sciences Discussions* **9**: 10447-10485.

Ballinger GA. 2004. Using generalized estimating equations for longitudinal data analysis *Organizational Research Methods* **7**: 127-150.

Barko JW, Smart RM. 1981. Comparative influences of light and temperature on the growth and metabolism of selected submersed freshwater macrophytes *Ecological Monographs* **51**: 219-235.

Barnett TP, Adam JC, Lettenmaier DP. 2005. Potential impacts of a warming climate on water availability in snow-dominated regions *Nature* **438**: 303-309.

Barrand NE, James TD, Murray T. 2010. Spatio-temporal variability in elevation changes of two high-Arctic valley glaciers *Journal of Glaciology* **56**: 771-780.

Bencala KE, Walters RA. 1983. Simulation of solute transport in a mountain pool-and-riffle stream: a transient storage model *Water Resources Research* **19**: 718-724.

Bernhardt ES, Likens GE. 2002. Dissolved organic carbon enrichment alters nitrogen dynamics in a forest stream *Ecology* **83**: 1689-1700.

Blaen PJ, Hannah DM, Brown LE, Milner AM. 2012. Water temperature dynamics in High Arctic river basins *Hydrological Processes*.

Brittain JE, Gislason GM, Bogen J, Olafsson J. 2009. Arctic Rivers. In *Rivers of Europe,* Tockner K, Uehlinger U, Robinson CT (eds). Academic Press.

Brosten TR, Bradford JH, Mcnamara JP, Gooseff MN, Zarnetske JP, Bowden WB, Johnston ME. 2009. Estimating 3D variation in active-layer thickness beneath Arctic streams using ground-penetrating radar *Journal of Hydrology* **373**: 479-486.

Brown GH. 2002. Glacier meltwater hydrochemistry *Applied Geochemistry* **17**: 855-883.

Brown JH, Gillooly JF, Allen AP, Savage VM, West GB. 2004. Toward a metabolic theory of ecology *Ecology* **85**: 1771-1789.

Brown L, Hannah D, Milner A. 2003. Alpine stream habitat classification: an alternative approach incorporating the role of dynamic water source contributions *Arctic, Antarctic, and Alpine Research* **35**: 313-322.

Brown LE, Hannah DM. 2007. Alpine Stream Temperature Response to Storm Events *Journal of Hydrometeorology* **8**: 952-967.

Brown LE, Hannah DM. 2008. Spatial heterogeneity of water temperature across an alpine river basin *Hydrological Processes* **22**: 954-967.

Brown LE, Hannah DM, Milner AM. 2005. Spatial and temporal water column and streambed temperature dynamics within an alpine catchment: implications for benthic communities *Hydrological Processes* **19**: 1585-1610.

Brown LE, Hannah DM, Milner AM. 2006a. Hydroclimatological influences on water column and streambed thermal dynamics in an alpine river system *Journal of Hydrology* **325**: 1-20.

Brown LE, Hannah DM, Milner AM. 2007a. Vulnerability of alpine stream biodiversity to shrinking glaciers and snowpacks *Global Change Biology* **13**: 958-966.

Brown LE, Milner AM, Hannah DM. 2007b. Groundwater influence on alpine stream ecosystems *Freshwater Biology* **52:** 878-890.

Brown LE, Hannah DM, Milner AM. 2009. ARISE: a classification tool for Alpine RIver and Stream Ecosystems *Freshwater Biology* **54**: 1357-1369.

Brown LE, Hannah DM, Milner AM, Soulsby C, Hodson AJ, Brewer MJ. 2006b. Water source dynamics in a glacierized alpine river basin (Taillon-Gabiétous, French Pyrénées) *Water Resources Research* **42**.

Brown LE, Milner AM. 2012. Rapid loss of glacial ice reveals stream community assembly processes *Global Change Biology*.

Brown S, Caesar J, Ferro C. 2008. Global changes in extreme daily temperature since 1950 *Journal of Geophysical Research* **113**: D05115.

Burgherr P, Ward J. 2001. Zoobenthos of kryal and lake outlet biotopes in a glacial flood plain *Internationale Vereinigung fur Theoretische und Angewandte Limnologie Verhandlungen* **27**: 1587-1590.

Cable J, Ogle K, Williams D. 2011. Contribution of glacier meltwater to streamflow in the Wind River Range, Wyoming, inferred via a Bayesian mixing model applied to isotopic measurements *Hydrological Processes* **25**: 2228-2236.

Cadbury SL, Hannah DM, Milner AM, Pearson CP, Brown LE. 2008. Stream temperature dynamics within a New Zealand glacierized river basin *River Research and Applications* **24**: 68-89.

Caine N. 1989. Hydrograph separation in a small alpine basin based on inorganic solute concentrations *Journal of Hydrology* **112**: 89-101.

Caissie D. 2006. The thermal regime of rivers: a review *Freshwater Biology* **51**: 1389-1406.

Callaghan T, Johansson M, Anisimov O, Christiansen H, Instanes A, Romanovsky V, Smith S. 2011a. Chapter 5: Changing permafrost and its impacts in *Snow, Water, Ice and Permafrost in the Arctic (SWIPA)*. Arctic Monitoring and Assessment Programme (AMAP): Oslo.

Callaghan T, Johansson M, Prowse T, Olsen M, Reiersen LO. 2011b. Arctic cryosphere: Changes and impacts *AMBIO: A Journal of the Human Environment* **40**: 3-5.

Callaghan TV, Johansson M, Brown RD, Groisman PY, Labba N, Radionov V, Barry RG, Bulygina ON, Essery RLH, Frolov D. 2011c. The changing face of Arctic snow cover: A synthesis of observed and projected changes *AMBIO: A Journal of the Human Environment* **40**: 17-31.

Campbell DH, Clow DW, Ingersoll GP, Mast MA, Spahr NE, Turk JT. 1995. Processes controlling the chemistry of two snowmelt-dominated streams in the Rocky Mountains *Water Resources Research* **31**: 2811-2821.

Cannone N, Guglielmin M, Gerdol R. 2004. Relationships between vegetation patterns and periglacial landforms in northwestern Svalbard *Polar Biology* **27:** 562-571.

Carey SK, Woo M. 2001. Spatial variability of hillslope water balance, Wolf Creek basin, subarctic Yukon *Hydrological Processes* **15**: 3113-3132.

Castella E, Adalsteinsson H, Brittain JE, Gislason GM, Lehmann A, Lencioni V, Lods-Crozet B, Maiolini B, Milner AM, Olafsson JS. 2001. Macrobenthic invertebrate richness and composition along a latitudinal gradient of European glacier-fed streams *Freshwater Biology* **46**: 1811-1831.

Cauwet G, Sidorov I. 1996. The biogeochemistry of Lena River: organic carbon and nutrients distribution *Marine Chemistry* **53**: 211-227.

Cavalli G, Riis T, Baattrup-Pedersen A. 2011. Bicarbonate use in three aquatic plants *Aquatic Botany* **98**: 57-60.

Chikita KA, Kaminaga R, Kudo I, Wada T, Kim Y. 2010. Parameters determining water temperature of a proglacial stream: the Phelan Creek and the Gulkana Glacier, Alaska *River Research and Applications* **26**: 995-1004.

Choiński A. 1989. Hydrology of the mouth section of the Ebbaelva and the Petuniabukta, Billefjorden, central Spitsbergen *Polish Polar Research* **10**: 457-464.

Christensen JH, Hewitson B, Busuioc A, Chen A, Gao X, Held R, Jones R, Kolli RK, Kwon W, Laprise R. 2007. Regional climate projections in *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change,* Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt K, Tignor M, Miller H (eds). Cambridge University Press: Cambridge.

Christophersen N, Hooper RP. 1992. Multivariate analysis of stream water chemical data: The use of principal components analysis for the end-member mixing problem *Water Resources Research* **28**: 99-107.

Cohen JL, Furtado JC, Barlow MA, Alexeev VA, Cherry JE. 2012. Arctic warming, increasing snow cover and widespread boreal winter cooling *Environmental Research Letters* **7**: 014007.

Cole J, Prairie Y, Caraco N, Mcdowell W, Tranvik L, Striegl R, Duarte C, Kortelainen P, Downing J, Middelburg J. 2007. Plumbing the global carbon cycle: integrating inland waters into the terrestrial carbon budget *Ecosystems* **10**: 172-185.

Comiso JC, Parkinson CL, Gersten R, Stock L. 2008. Accelerated decline in the Arctic sea ice cover *Geophysical Research Letters* **35**: L01703.

Cooper EJ, Alsos IG, Hagen D, Smith FM, Coulson SJ, Hodkinson ID. 2004. Plant recruitment in the High Arctic: seed bank and seedling emergence on Svalbard *Journal of Vegetation Science* **15**: 115-124.

Cooper R, Hodgkins R, Wadham J, Tranter M. 2011. The hydrology of the proglacial zone of a high-Arctic glacier (Finsterwalderbreen, Svalbard): Sub-surface water fluxes and complete water budget *Journal of Hydrology* **406**: 88-96.

Cooper R, Wadham J, Tranter M, Hodgkins R, Peters N. 2002. Groundwater hydrochemistry in the active layer of the proglacial zone, Finsterwalderbreen, Svalbard *Journal of Hydrology* **269**: 208-223.

Coulson S, Hodkinson I, Webb N, Mikkola K, Harrison J, Pedgley D. 2002. Aerial colonization of high Arctic islands by invertebrates: the diamondback moth Plutella xylostella (Lepidoptera: Yponomeutidae) as a potential indicator species *Diversity and Distributions* **8**: 327-334.

Coulson SJ. 2000. A review of the terrestrial and freshwater invertebrate fauna of the High Arctic archipelago of Svalbard *Norwegian Journal of Entomology* **47**: 41-63.

Coulson SJ. 2007. The terrestrial and freshwater invertebrate fauna of the High Arctic archipelago of Svalbard *Zootaxa* **1448**: 41-58.

Covino TP, Mcglynn BL, Mcnamara RA. 2010. Tracer Additions for Spiraling Curve Characterization (TASCC): Quantifying stream nutrient uptake kinetics from ambient to saturation *Limnology and Oceanography: Methods* **8**: 484-498.

Cozzetto K, McKnight D, Nylen T, Fountain A. 2006. Experimental investigations into processes controlling stream and hyporheic temperatures, Fryxell Basin, Antarctica *Advances in Water Resources* **29:** 130-153.

Craig LS, Palmer MA, Richardson DC, Filoso S, Bernhardt ES, Bledsoe BP, Doyle MW, Groffman PM, Hassett BA, Kaushal SS. 2008. Stream restoration strategies for reducing river nitrogen loads *Frontiers in Ecology and the Environment* **6**: 529-538.

Craig P, McCart P, 1975. Classification of stream types in Beaufort Sea drainages between Prudhoe Bay, Alaska, and the Mackenzie delta, NWT, Canada *Arctic and Alpine Research* **7:** 183-198.

Crossman J, Bradley C, Boomer I, Milner AM. 2011. Water Flow Dynamics of Groundwater-Fed Streams and Their Ecological Significance in a Glacierized Catchment *Arctic, Antarctic, and Alpine Research* **43**: 364-379.

Dahlke HE, Lyon SW, Stedinger JR, Rosqvist G, Jansson P. 2012. Contrasting trends in hydrologic extremes for two sub-Arctic catchments in northern Sweden; does glacier melt matter? *Hydrology and Earth System Sciences Discussions* **9**: 1041-1084.

Dallmann WK, Ohta Y., Elvevold S., D. B. 2002. *Bedrock map of Svalbard and Jan Mayen: Norsk Polarinstitutt Thematic Map, no. 33*. Norsk polarinstitutt: Norway.

Danks H. 2007. How aquatic insects live in cold climates *The Canadian Entomologist* **139**: 443-471.

Demars BOL, Russell Manson J, Ólafsson JS, Gíslason GM, Gudmundsdóttir R, Woodward G, Reiss J, Pichler DE, Rasmussen JJ, Friberg N. 2011. Temperature and the metabolic balance of streams *Freshwater Biology* **56**: 1106-1121.

Deutsch CA, Tewksbury JJ, Huey RB, Sheldon KS, Ghalambor CK, Haak DC, Martin PR. 2008. Impacts of climate warming on terrestrial ectotherms across latitude *Proceedings of the National Academy of Sciences* **105**: 6668-6672.

Dittmar T, Kattner G. 2003. The biogeochemistry of the river and shelf ecosystem of the Arctic Ocean: A review *Marine Chemistry* **83**: 103-120.

Dodds WK, Marti E, Tank JL, Pontius J, Hamilton SK, Grimm NB, Bowden WB, Mcdowell WH, Peterson BJ, Valett HM. 2004. Carbon and nitrogen stoichiometry and nitrogen cycling rates in streams *Oecologia* **140**: 458-467.

Dragon K, Marciniak M. 2010a. Chemical composition of groundwater and surface water in the Arctic environment (Petuniabukta region, central Spitsbergen) *Journal of Hydrology* **386**: 160-172.

Dragon K, Marciniak M. 2010b. Chemical composition of groundwater and surface water in the Arctic environment (Petuniabukta region, central Spitsbergen) *Journal of Hydrology* **386**: 160-172.

Dugan HA, Lamoureux SF, Lafrenière MJ, Lewis T. 2009. Hydrological and sediment yield response to summer rainfall in a small high Arctic watershed *Hydrological Processes* **23**: 1514-1526.

Durand J, Lusardi R, Nover D, Suddeth R, Carmona-Catot G, Connell-Buck C, Gatzke S, Katz J, Mount J, Moyle P. 2011. Environmental heterogeneity and community structure of the Kobuk River, Alaska, in response to climate change *Ecosphere* **2(4):** 1-19.

Dzikowski M, Jobard S. 2011. Mixing law versus discharge and electrical conductivity relationships: application to an alpine proglacial stream *Hydrological Processes* **26**: 2724-2732.

Edwardson KJ, Bowden WB, Dahm C, Morrice J. 2003. The hydraulic characteristics and geochemistry of hyporheic and parafluvial zones in Arctic tundra streams, north slope, Alaska *Advances in Water Resources* **26**: 907-923.

eKlima. 2012. *Free access to weather and climate data from Norwegian Meteorological Institute from historical data to time observations* Available at: http://www.eklima.no Accessed: 05/02/2012

Engeset RV, Weydahl DJ. 1998. Analysis of glaciers and geomorphology on Svalbard using multitemporal ERS-1 SAR images *Geoscience and Remote Sensing, IEEE Transactions on* **36**: 1879-1887.

Ensign SH, Doyle MW. 2006. Nutrient spiraling in streams and river networks *Journal of Geophysical Research* **111:** G04009.

Etzelmüller B, Schuler T, Isaksen K, Christiansen H, Farbrot H, Benestad R. 2011. Modeling the temperature evolution of Svalbard permafrost during the 20th and 21st century *The Cryosphere* **5**: 67-79.

Evans E, Mcgregor GR, Petts GE. 1998. River energy budgets with special reference to river bed processes *Hydrological Processes* **12**: 575-595.

Evengard B, Berner J, Brubaker M, Mulvad G, Revich B. 2011. Climate change and water security with a focus on the Arctic *Global Health Action* **4**: 8449-8453.

Finn DS, Khamis K, Milner AM. 2012. Loss of small glaciers will diminish beta diversity in Pyrenean streams at two levels of biological organization *Global Ecology and Biogeography* **22**: 40-51.

Finn DS, Poff NL. 2005. Variability and convergence in benthic communities along the longitudinal gradients of four physically similar Rocky Mountain streams *Freshwater Biology* **50**: 243-261.

Fleming SW, Clarke GKC. 2003. Glacial control of water resource and related environmental responses to climatic warming: empirical analysis using historical streamflow data from northwestern Canada *Canadian Water Resources Journal* **28**: 69-86.

Fleming SW, Clarke GK. 2006. Glacier-mediated streamflow teleconnections to the Arctic Oscillation *International Journal of Climatology* **26**: 619-636.

Førland EJ, Benestad R, Hanssen-Bauer I, Haugen JE, Skaugen TE. 2011. Temperature and Precipitation Development at Svalbard 1900–2100 *Advances in Meteorology* **2011:** 1-14.

Førland EJ, Hanssen-Bauer I. 2003. Past and future climate variations in the Norwegian Arctic: overview and novel analyses *Polar Research* **22**: 113-124.

Frey KE, Mcclelland JW. 2009. Impacts of permafrost degradation on arctic river biogeochemistry *Hydrological Processes* **23**: 169-182.

Frey KE, Smith LC. 2005. Amplified carbon release from vast West Siberian peatlands by 2100 *Geophysical Research Letters* **32**: L09401.

Friberg N, Bergfur J, Rasmussen J, Sandin L. 2013. Changing Northern catchments: Is altered hydrology, temperature or both going to shape future stream communities and ecosystem processes? *Hydrological Processes*.

Friberg N, Dybkjaer JB, Olafsson JS, Gislason GM, Larsen SE, Lauridsen TL. 2009. Relationships between structure and function in streams contrasting in temperature *Freshwater Biology* **54**: 2051-2068.

Friberg N, Milner AM, Svendsen LM, Lindegaard C, Larsen SE. 2001. Macroinvertebrate stream communities along regional and physico-chemical gradients in Western Greenland *Freshwater Biology* **46**: 1753-1764.

Fritioff A, Kautsky L, Greger M. 2005. Influence of temperature and salinity on heavy metal uptake by submersed plants *Environmental Pollution* **133**: 265-274.

Füreder L. 2007. Life at the Edge: Habitat Condition and Bottom Fauna of Alpine Running Waters *International Review of Hydrobiology* **92**: 491-513.

Füreder L, Schütz C, Wallinger M, Burger R. 2001. Physico-chemistry and aquatic insects of a glacier-fed and a spring-fed alpine stream *Freshwater Biology* **46**: 1673-1690.

Genereux D. 1998. Quantifying uncertainty in tracer-based hydrograph separations *Water Resources Research* **34**: 915-919.

Gersper P, Alexander V, Barkley S, Barsdate R, Flint P. 1980. The soils and their nutrients In *An Arctic ecosystem: the coastal tundra at Barrow, Alaska*. Brown J, Miller PC, Tieszen L, Bunnell, FD (eds). Hutchison and Ross: Pennsylvania, USA: 219-254.

Gibson J. 2002. Short-term evaporation and water budget comparisons in shallow Arctic lakes using non-steady isotope mass balance *Journal of Hydrology* **264**: 242-261.

Graversen RG, Mauritsen T, Tjernström M, Källén E, Svensson G. 2008. Vertical structure of recent Arctic warming *Nature* **451**: 53-56.

Greenwald MJ, Bowden WB, Gooseff MN, Zarnetske JP, Mcnamara JP, Bradford JH, Brosten TR. 2008. Hyporheic exchange and water chemistry of two Arctic tundra streams of contrasting geomorphology *Journal of Geophysical Research* **113**.

Gudmundsdottir R, Olafsson J, Palsson S, Gislason GM, Moss B. 2011. How will increased temperature and nutrient enrichment affect primary producers in sub-Arctic streams? *Freshwater Biology* **56**: 2045-2058.

Gücker B, Boëchat IG. 2004. Stream morphology controls ammonium retention in tropical headwaters *Ecology* **85**: 2818-2827.

Gurnell A, Clark M, Hill C, Greenhalgh J. 1992. Reliability and representativeness of a suspended sediment concentration monitoring programme for a remote alpine proglacial river In *Proceedings of the Oslo Symposium on Erosion and Sediment Transport Monitoring in River Basins*. Oslo: 28.

Haeberli W, Hoelzle M, Zemp M. 2007. *Glacier mass balance bulletin No. 9 (2004-05)*. World Glacier Monitoring Service: Zurich.

Hagen JO, Lefauconnier B. 1995. Reconstructed runoff from the high Arctic basin Bayelva based on mass-balance measurements *Nordic Hydrology* **26**: 285-296.

Haggard B, Storm D, Tejral R, Popova Y, Keyworth V, Stanley E. 2001. Stream nutrient retention in three northeastern Oklahoma agricultural catchments *Transactions of the American Society of Agricultural Engineers* **44**: 597-605.

Haldorsen S, Heim M, Dale B, Landvik JY, Van Der Ploeg M, Leijnse A, Salvigsen O, Hagen JO, Banks D. 2010. Sensitivity to long-term climate change of subpermafrost groundwater systems in Svalbard *Quaternary Research* **73**: 393-402.

Haldorsen S, Heim M, Lefauconnier B, Pettersson LE, Røros M, Sandsbråten K. 2002. The water balance of an arctic lake and its dependence on climate change: Tvillingvatnet in Ny-Ålesund, Svalbard *Norsk Geografisk Tidsskrift* **56**: 146-151.

Halekoh U, Højsgaard S, Yan J. 2006. The R package geepack for generalized estimating equations *Journal of Statistical Software* **15**: 1-11.

Hall RO, Bernhardt ES, Likens GE. 2002. Relating nutrient uptake with transient storage in forested mountain streams *Limnology and Oceanography* **47**: 255-265.

Hall RO, Tank JL. 2003. Ecosystem metabolism controls nitrogen uptake in streams in Grand Teton National Park, Wyoming *Limnology and Oceanography* **48**: 1120-1128.

Hallet B, Prestrud S. 1986. Dynamics of periglacial sorted circles in western Spitsbergen *Quaternary Research* **26**: 81-99.

Hambrey MJ, Murray T, Glasser NF, Hubbard A, Hubbard B, Stuart G, Hansen S, Kohler J. 2005. Structure and changing dynamics of a polythermal valley glacier on a centennial timescale: Midre Lovenbreen, Svalbard *Journal of Geophysical Research* **110**: F01006.

Han L. 1997. Spectral reflectance with varying suspended sediment concentrations in clear and algae-laden waters *Photogrammetric Engineering and Remote Sensing* **63**: 701-705.

Hannah DM, Brown LE, Milner AM, Gurnell AM, Mcgregor GR, Petts GE, Smith BPG, Snook DL. 2007. Integrating climate–hydrology–ecology for alpine river systems *Aquatic Conservation: Marine and Freshwater Ecosystems* **17**: 636-656.

Hannah DM, Malcolm IA, Bradley C. 2009. Seasonal hyporheic temperature dynamics over riffle bedforms *Hydrological Processes* **23**: 2178-2194.

Hannah DM, Malcolm IA, Soulsby C, Youngson AF. 2004. Heat exchanges and temperatures within a salmon spawning stream in the Cairngorms, Scotland: seasonal and sub-seasonal dynamics *River Research and Applications* **20**: 635-652.

Hannah DM, Malcolm IA, Soulsby C, Youngson AF. 2008. A comparison of forest and moorland stream microclimate, heat exchanges and thermal dynamics *Hydrological Processes* **22**: 919-940.

Hardin JW, Hilbe JM. 2003. *Generalized Estimating Equations (GEE)*. Chapman and Hall: Boca Raton, FL.

Harland WB. 1997. *The Geology of Svalbard*. Geological Society Publishing House: Bath.

Harms TK, Jones, JB. 2012. Thaw depth determines reaction and transport of inorganic nitrogen in valley bottom permafrost soils *Global Change Biology* **18:** 2958-2968

Hebert C, Caissie D, Satish MG, El-Jabi N. 2011. Study of stream temperature dynamics and corresponding heat fluxes within Miramichi River catchments (New Brunswick, Canada) *Hydrological Processes* **25**: 2439-2455.

Hershey A, Bowden W, Deegan L, Hobbie J, Peterson B, Kipphut G, Kling G, Lock M, Merritt R, Miller M. 1997. The Kuparuk River: a long-term study of biological and chemical processes in an Arctic river in *Freshwaters of Alaska*, Milner AM, Oswood MW (eds). Springer-Verlag: New York

Hieber M, Robinson CT, Uehlinger U, Ward J. 2002. Are alpine lake outlets less harsh than other alpine streams? *Archiv für Hydrobiologie* **154**: 199-223.

Hinton M, Schiff S, English M. 1994. Examining the contributions of glacial till water to storm runoff using two- and three-component hydrograph separation *Water Resources Research* **30**: 983-994.

Hinzman L, Kane D, Gieck R, Everett K. 1991. Hydrologic and thermal properties of the active layer in the Alaskan Arctic *Cold Regions Science and Technology* **19**: 95-110.

Hinzman LD, Bettez ND, Bolton WR, Chapin FS, Dyurgerov MB, Fastie CL, Griffith B, Hollister RD, Hope A, Huntington HP, Jensen AM, Jia GJ, Jorgenson T, Kane DL, Klein DR, Kofinas G, Lynch AH, Lloyd AH, Mcguire AD, Nelson FE, Oechel WC, Osterkamp TE, Racine CH, Romanovsky VE, Stone RS, Stow DA, Sturm M, Tweedie CE, Vourlitis GL, Walker MD, Walker DA, Webber PJ, Welker JM, Winker KS, Yoshikawa K. 2005. Evidence and Implications of Recent Climate Change in Northern Alaska and Other Arctic Regions *Climatic Change* **72**: 251-298.

Hjelle A. 1993. *Geology of Svalbard*. Norsk Polarinstitutt: Tromso.

Hock R. 2005. Glacier melt: a review of processes and their modelling *Progress in Physical Geography* **29**: 362-391.

Hodgkins R. 1997. Glacier hydrology in Svalbard, Norwegian high arctic *Quaternary Science Reviews* **16**: 957-973.

Hodgkins R. 2001. Seasonal evolution of meltwater generation, storage and discharge at a non-temperate glacier in Svalbard *Hydrological Processes* **15**: 441-460.

Hodgkins R, Cooper R, Wadham J, Tranter M. 2009. The hydrology of the proglacial zone of a high-Arctic glacier (Finsterwalderbreen, Svalbard): Atmospheric and surface water fluxes *Journal of Hydrology* **378**: 150-160.

Hodson A, Gurnell A, Tranter M, Bogen J, Hagen JO, Clark M. 1998a. Suspended sediment yield and transfer processes in a small High-Arctic glacier basin, Svalbard *Hydrological Processes* **12**: 73-86.

Hodson A, Gurnell A, Washington R, Tranter M, Clark M, Hagen J. 1998b. Meteorological and runoff time-series characteristics in a small, high-Arctic glaciated basin, Svalbard *Hydrological Processes* **12**: 509-526.

Hodson A, Roberts TJ, Engvall AC, Holmén K, Mumford P. 2010. Glacier ecosystem response to episodic nitrogen enrichment in Svalbard, European High Arctic *Biogeochemistry* **98**: 171-184.

Hodson A, Tranter M, Gurnell A, Clark M, Hagen JO. 2002. The hydrochemistry of Bayelva, a high Arctic proglacial stream in Svalbard *Journal of Hydrology* **257**: 91-114.

Hodson A, Tranter M, Vatne G. 2000. Contemporary rates of chemical denudation and atmospheric CO_2 sequestration in glacier basins: an Arctic perspective *Earth Surface Processes and Landforms* **25**: 1447-1471.

Hodson AJ, Mumford PN, Kohler J, Wynn PM. 2005. The High Arctic glacial ecosystem: new insights from nutrient budgets *Biogeochemistry* **72**: 233-256.

Hoellein TJ, Tank JL, Rosi-Marshall EJ, Entrekin SA, Lamberti GA. 2007. Controls on spatial and temporal variation of nutrient uptake in three Michigan headwater streams *Limnology and Oceanography* **52**: 1964-1977.

Høj L, Rusten M, Haugen LE, Olsen RA, Torsvik VL. 2006. Effects of water regime on archaeal community composition in Arctic soils *Environmental Microbiology* **8**: 984-996.

Holland HD. 1978. *The chemistry of the atmosphere and oceans*. Wiley: New York.

Holland MM, Bitz CM. 2003. Polar amplification of climate change in coupled models *Climate Dynamics* **21**: 221-232.

Holmes RM, Mcclelland JW, Peterson BJ, Tank SE, Bulygina E, Eglinton TI, Gordeev VV, Gurtovaya TY, Raymond PA, Repeta DJ. 2012. Seasonal and annual fluxes of nutrients and organic matter from large rivers to the Arctic Ocean and surrounding seas *Estuaries and Coasts* **35**: 369-382.

Holmes RM, Mcclelland JW, Raymond PA, Frazer BB, Peterson BJ, Stieglitz M. 2008. Lability of DOC transported by Alaskan rivers to the Arctic Ocean *Geophysical Research Letters* **35**: L03402.

Hood E, Berner L. 2009. Effects of changing glacial coverage on the physical and biogeochemical properties of coastal streams in southeastern Alaska *Journal of Geophysical Research* **114**: G03001.

Hood E, Fellman J, Spencer RGM, Hernes PJ, Edwards R, D'amore D, Scott D. 2009. Glaciers as a source of ancient and labile organic matter to the marine environment *Nature* **462**: 1044-1047.

Howell T, Meek D, Hatfield J. 1983. Relationship of photosynthetically active radiation to shortwave radiation in the San Joaquin Valley *Agricultural Meteorology* **28**: 157-175.

Hudson R, Fraser J. 2005. The Mass Balance (or Dry Injection) Method *Streamline Watershed Management Bulletin* **9**: 6-12.

Huryn AD, Slavik KA, Lowe RL, Parker SM, Anderson DS, Peterson BJ. 2005. Landscape heterogeneity and the biodiversity of Arctic stream communities: a habitat template analysis *Canadian Journal of Fisheries and Aquatic Sciences* **62**: 1905-1919.

Ilg C, Castella E. 2006. Patterns of macroinvertebrate traits along three glacial stream continuums *Freshwater Biology* **51**: 840-853.

Ilg C, Castella E, Lods-Crozet B, Marmonier P. 2001. Invertebrate drift and physico-chemical variables in the tributaries of the Mutt, a Swiss glacial stream *Archiv für Hydrobiologie* **151**: 335-352.

Irons JG, Oswood MW. 1992. Seasonal temperature patterns in an arctic and two subarctic Alaskan (USA) headwater streams *Hydrobiologia* **237**: 147-157.

Irvine-Fynn T, Barrand N, Porter P, Hodson A, Murray T. 2011. Recent High-Arctic glacial sediment redistribution: A process perspective using airborne lidar *Geomorphology* **125**: 27-39.

Isaak DJ, Hubert WA. 2001. A hypothesis about factors that affect maximum summer stream temperatures across montane landscapes *Journal of the American Water Resources Association* **37**: 351-366.

Jacobsen D, Milner AM, Brown LE, Dangles O. 2012. Biodiversity under threat in glacier-fed river systems *Nature Climate Change* **2:** 361-364.

James T, Murray T, Barrand N, Sykes H, Fox A, King M. 2012. Observations of enhanced thinning in the upper reaches of Svalbard glaciers *The Cryosphere* **6**: 1369-1381.

Johnson LT, Tank JL, Arango CP. 2009. The effect of land use on dissolved organic carbon and nitrogen uptake in streams *Freshwater Biology* **54**: 2335-2350.

Joly D, Brossard T. 2007. Contribution of environmental factors to temperature distribution at different resolution levels on the forefield of the Loven Glaciers, Svalbard *Polar Record* **43**: 353-360.

Jones JB, Petrone KC, Finlay JC, Hinzman LD, Bolton WR. 2005. Nitrogen loss from watersheds of interior Alaska underlain with discontinuous permafrost *Geophysical Research Letters* **32**: L02401.

Jones JB, Rinehart AJ. 2010. The long-term response of stream flow to climatic warming in headwater streams of interior Alaska *Canadian Journal of Forest Research* **40**: 1210-1218.

Judd KE, Kling GW. 2002. Production and export of dissolved C in arctic tundra mesocosms: the roles of vegetation and water flow *Biogeochemistry* **60**: 213-234.

Juggins S. 2007. C2 Version 1.5 User guide. Software for ecological and palaeoecological data analysis and visualisation. Newcastle University, Newcastle upon Tyne, UK.

Kaplan JO, New M. 2006. Arctic climate change with a 2 ºC global warming: Timing, climate patterns and vegetation change *Climatic Change* **79**: 213-241.

Kattsov VM, Källén E, Cattle H, Christensen J, Drange H, Hanssen-Bauer I, Jóhannesen T, Karol I, Räisänen J, Svensson G, Vavulin S. 2005. Chapter 4: Future Climate Change: Modeling and Scenarios in *Arctic Climate Impact Assessment*. Cambridge University Press: Cambridge.

Kattsov VM, Walsh JE, Chapman WL, Govorkova VA, Pavlova TV, Zhang X. 2007. Simulation and projection of Arctic freshwater budget components by the IPCC AR4 global climate models *Journal of Hydrometeorology* **8**: 571-589.

Kaufman DS, Schneider DP, Mckay NP, Ammann CM, Bradley RS, Briffa KR, Miller GH, Otto-Bliesner BL, Overpeck JT, Vinther BM. 2009. Recent warming reverses long-term Arctic cooling *Science* **325**: 1236-1239.

Kelley AM, Epstein HE. 2009. Effects of nitrogen fertilization on plant communities of nonsorted circles in moist nonacidic tundra, northern Alaska *Arctic, Antarctic, and Alpine Research* **41**: 119-127.

Keuper F, Bodegom PM, Dorrepaal E, Weedon JT, Hal J, Logtestijn RSP, Aerts R. 2012. A frozen feast: thawing permafrost increases plant-available nitrogen in subarctic peatlands *Global Change Biology* **18:** 1998-2007

Killingtveit Å, Pettersson LE, Sand K. 2003. Water balance investigations in Svalbard *Polar Research* **22**: 161-174.

Kim GH, Klochkova TA, Kang SH. 2008. Notes on freshwater and terrestrial algae from Ny-Ålesund, Svalbard (High Arctic sea area) *Journal of Environmental Biology* **29**: 485-491.

Knispel S, Castella E. 2003. Disruption of a longitudinal pattern in environmental factors and benthic fauna by a glacial tributary *Freshwater Biology* **48**: 604-618.

Kobayashi D, Ishii Y, Kodama Y. 1999. Stream temperature, specific conductance and runoff process in mountain watersheds *Hydrological Processes* **13**: 865-876.

Kostrzewski A, Kaniecki A, Kapuscinski J, Klimczak R, Stach A, Zwolinski Z. 1989. The dynamics and rate of denudation of glaciated and non-glaciated catchments, central Spitsbergen *Polish Polar Research* **10**: 317-367.

Krawczyk WE, Bartoszewski SA. 2008. Crustal solute fluxes and transient carbon dioxide drawdown in the Scottbreen Basin, Svalbard in 2002 *Journal of Hydrology* **362**: 206-219.

Kühnel R, Roberts TJ, Björkman MP, Isaksson E, Aas W, Holmén K, Ström J. 2011. 20-Year climatology of NO_3^- and NH_4^+ wet deposition at Ny-Alesund, Svalbard *Advances in Meteorology* **10**: 1-10.

Lamb H, Tranter M, Brown G, Hubbard B, Sharp M, Gordon S, Smart C, Willis I, Nielsen M. 1995. The composition of subglacial meltwaters sampled from boreholes at the Haut Glacier d'Arolla, Switzerland *IAHS Publications Series of Proceedings and Reports* **228**: 395-404.

Lammers RB, Pundsack JW, Shiklomanov AI. 2007. Variability in river temperature, discharge, and energy flux from the Russian pan-Arctic landmass *Journal of Geophysical Research* **112**: G04S59.

Lappegard G, Beldring S, Roald L, Engen-Skaugen T, Førland E. 2007. Projection of future streamflow in glaciated and non-glaciated catchments in Norway. *Norwegian Water Resources and Energy Directorate* Consultancy Report A (9): 64.

Laudon H, Slaymaker O. 1997. Hydrograph separation using stable isotopes, silica and electrical conductivity: an alpine example *Journal of Hydrology* **201**: 82-101.

Lawrence DM, Slater AG. 2005. A projection of severe near-surface permafrost degradation during the 21st century *Geophysical Research Letters* **32**: L24401.

Lecerf A, Richardson J. 2010. Biodiversity-ecosystem function research: Insights gained from streams *River Research and Applications* **26**: 45-54.

Lee J, Feng X, Faiia AM, Posmentier ES, Kirchner JW, Osterhuber R, Taylor S. 2010. Isotopic evolution of a seasonal snowcover and its melt by isotopic exchange between liquid water and ice *Chemical Geology* **270**: 126-134.

Legendre P, Gallagher ED. 2001. Ecologically meaningful transformations for ordination of species data *Oecologia* **129**: 271-280.

Lemke P, Ren J, Alley RB, Allison I, Carrasco J, Flato G, Fujii Y, Kaser G, Mote P, Thomas RH. 2007. Observations: Changes in snow, ice and frozen ground in *Climate Change 2007: the physical science basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL (eds.). Cambridge University Press: Cambridge.

Lencioni V. 2004. Survival strategies of freshwater insects in cold environments *Journal of Limnology* **63**: 45-55.

Lischeid G, Kolb A, Alewell C. 2002. Apparent translatory flow in groundwater recharge and runoff generation *Journal of Hydrology* **265**: 195-211.

Liu F, Williams MW, Caine N. 2004. Source waters and flow paths in an alpine catchment, Colorado Front Range, United States *Water Resources Research* **40**: W09401.

Lobbes JM, Fitznar HP, Kattner G. 2000. Biogeochemical characteristics of dissolved and particulate organic matter in Russian rivers entering the Arctic Ocean *Geochimica et Cosmochimica Acta* **64**: 2973-2983.

Lock MA, Ford TE, Hullar MaJ, Kaufman M, Robie Vestal J, Volk GS, Ventullo RM. 1990. Phosphorous limitation in an Arctic river biofilm—a whole ecosystem experiment *Water Research* **24**: 1545-1549.

Lods-Crozet B, Lencioni V, Olafsson J, Snook D, Velle G, Brittain J, Castella E, Rossaro B. 2001. Chironomid (Diptera: Chironomidae) communities in six European glacier-fed streams *Freshwater Biology* **46**: 1791-1809.

Lods-Crozet B, Lencioni V, Brittain JE, Marziali L, Rossaro B. 2007. Contrasting chironomid assemblages in two high Arctic streams on Svalbard *Fundamental and Applied Limnology/Archiv für Hydrobiologie* **170:** 211-222.

Logue JB, Robinson CT, Meier C, Van Der Meer JR. 2004. Relationship between sediment organic matter, bacteria composition, and the ecosystem metabolism of alpine streams *Limnology and Oceanography* **49**: 2001-2010.

Lu J, Cai M. 2009. Seasonality of polar surface warming amplification in climate simulations *Geophysical Research Letters* **36**: L16704.

Madan NJ, Deacon LJ, Robinson CH. 2007. Greater nitrogen and/or phosphorus availability increase plant species‘ cover and diversity at a High Arctic polar semidesert *Polar Biology* **30**: 559-570.

Mahlstein I, Portmann RW, Daniel JS, Solomon S, Knutti R. 2012. Perceptible changes in regional precipitation in a future climate *Geophysical Research Letters* **39**: L05701.

Malard F, Galassi D, Lafont M, Dolédec S, Ward J. 2003. Longitudinal patterns of invertebrates in the hyporheic zone of a glacial river *Freshwater Biology* **48**: 1709-1725.

Malard F, Tockner K, Ward J. 1999. Shifting dominance of subcatchment water sources and flow paths in a glacial floodplain, Val Roseg, Switzerland *Arctic, Antarctic, and Alpine Research*: 135-150.

Malard F, Tockner K, Ward J. 2000. Physico-chemical heterogeneity in a glacial riverscape *Landscape Ecology* **15**: 679-695.

Marcogliese D. 2008. The impact of climate change on the parasites and infectious diseases of aquatic animals *Revue scientifique et technique (International Office of Epizootics)* **27**: 467.

Marsh P, Woo MK. 1981. Snowmelt, glacier melt, and high arctic streamflow regimes *Canadian Journal of Earth Sciences* **18**: 1380-1384.

Martí E, Fonollà P, Schiller DV, Sabater F, Argerich A, Ribot M, Riera JL. 2009. Variation in stream C, N and P uptake along an altitudinal gradient: a space-for-time analogue to assess potential impacts of climate change *Hydrology Research* **40**: 123.

Marti E, Sabater F. 1996. High variability in temporal and spatial nutrient retention in Mediterranean streams *Ecology* **77**: 854-869.

Marziali L, Gozzini M, Rossaro B, Lencioni V. 2009. Drift patterns of Chironomidae (Insecta, Diptera) in an arctic stream (Svalbard Islands): an experimental approach *Studi Trentini di Scienze Naturali* **84**: 87-96.

McBean G, Alekseev G, Chen D, Foerland E, Fyfe J, Groisman PY, King R, Melling H, Vose R, Whitfield PH. 2005. Chapter 2: Arctic Climate: past and present in *Arctic Climate Impact Assessment*. Cambridge University Press: Cambridge.

McClelland JW, Holmes RM, Peterson BJ, Stieglitz M. 2004. Increasing river discharge in the Eurasian Arctic: Consideration of dams, permafrost thaw, and fires as potential agents of change *Journal of Geophysical Research* **109**: D18102.

McGuire AD, Melillo JM, Kicklighter DW, Joyce LA. 1995. Equilibrium responses of soil carbon to climate change: empirical and process-based estimates *Journal of Biogeography* **22:** 785-796.

McKnight DM., Smith RL, Harnish RA, Miller CL, Bencala KE. 1993. Seasonal relationships between planktonic microorganisms and dissolved organic material in an alpine stream *Biogeochemistry* **21:** 39-59.

McNamara JP, Kane DL, Hinzman LD. 1997. Hydrograph separations in an Arctic watershed using mixing model and graphical techniques *Water Resources Research* **33**: 1707-1719.

Merck M, Neilson B, Cory R, Kling G. 2012. Variability of in-stream and riparian storage in a beaded Arctic stream *Hydrological Processes*.

Meyer JL, Strayer DL, Wallace JB, Eggert SL, Helfman GS, Leonard NE. 2007. The Contribution of Headwater Streams to Biodiversity in River Networks *Journal of the American Water Resources Association* **43**: 86-103.

Milner AM, Brittain JE, Castella E, Petts GE. 2001. Trends of macroinvertebrate community structure in glacier-fed rivers in relation to environmental conditions: a synthesis *Freshwater Biology* **46**: 1833-1847.

Milner AM, Brown LE, Hannah DM. 2009. Hydroecological response of river systems to shrinking glaciers *Hydrological Processes* **23**: 62-77.

Milner AM, Robertson AL, Monaghan KA, Veal AJ, Flory EA. 2008. Colonization and development of an Alaskan stream community over 28 years *Frontiers in Ecology and the Environment* **6**: 413-419.

Mindl B, Anesio AM, Meirer K, Hodson AJ, Laybourn-Parry J, Sommaruga R, Sattler B. 2007. Factors influencing bacterial dynamics along a transect from supraglacial runoff to proglacial lakes of a high Arctic glacier *FEMS Microbiology Ecology* **59**: 307-317.

Mohseni O, Stefan H. 1999. Stream temperature/air temperature relationship: a physical interpretation *Journal of Hydrology* **218**: 128-141.

Moore RD, Fleming SW, Menounos B, Wheate R, Fountain A, Stahl K, Holm K, Jakob M. 2009. Glacier change in western North America: influences on hydrology, geomorphic hazards and water quality *Hydrological Processes* **23**: 42-61.

Morrice JA, Valett H, Dahm CN, Campana ME. 1997. Alluvial characteristics, groundwater–surface water exchange and hydrological retention in headwater streams *Hydrological Processes* **11**: 253-267.

Mulholland PJ, Tank J, Webster J, Bowden W, Dodds W, Gregory S, Grimm N, Hamilton S, Johnson S, Martí E. 2002. Can uptake length in streams be determined by nutrient addition experiments? Results from an interbiome comparison study *Journal of the North American Benthological Society* **21**: 544-560.

Mulholland PJ, Marzolf ER, Webster JR, Hart DR, Hendricks SP. 1997. Evidence that hyporheic zones increase heterotrophic metabolism and phosphorus uptake in forest streams *Limnology and Oceanography* **42**: 443-451.

Muraoka H, Noda H, Uchida M, Ohtsuka T, Koizumi H, Nakatsubo T. 2008. Photosynthetic characteristics and biomass distribution of the dominant vascular plant species in a High Arctic tundra ecosystem, Ny-Ålesund, Svalbard: implications for their role in ecosystem carbon gain *Journal of Plant Research* **121**: 137-145.

Nakatsubo T, Yoshitake S, Uchida M, Uchida M, Shibata Y, Koizumi H. 2008. Organic carbon and microbial biomass in a raised beach deposit under terrestrial vegetation in the High Arctic, Ny-Ålesund, Svalbard *Polar Research* **27**: 23-27.

Nelson F, Anisimov O, Shiklomanov NI. 2002. Climate change and hazard zonation in the circum-Arctic permafrost regions *Natural Hazards* **26**: 203-225.

Newbold JD, Bott TL, Kaplan LA, Dow CL, Jackson JK, Aufdenkampe AK, Martin LA, Van Horn DJ, Long AA. 2006. Uptake of nutrients and organic C in streams in New York City drinking-water-supply watersheds *Journal of the North American Benthological Society* **25**: 998-1017.

Obradovic MM, Sklash MG. 1986. An isotopic and geochemical study of snowmelt runoff in a small arctic watershed *Hydrological Processes* **1**: 15-30.

O'Donnell JA, Aiken GR, Kane ES, Jones JB. 2010. Source water controls on the character and origin of dissolved organic matter in streams of the Yukon River basin, Alaska *Journal of Geophysical Research* **115**: G03025.

Oke TR. 1987. *Boundary layer climates*. Routledge: UK.

Oksanen J, Kindt R, Legendre P, O‘hara B, Stevens MHH, Oksanen MJ, Suggests M. 2007. The vegan package *Community ecology package for R* **10**: 2007.

Oliver DR, Roussel ME. 1983. *The insects and arachnids of Canada, Part 11. The genera of larval midges of Canada. Diptera: Chironomidae*. Agriculture Canada: Canada.

Ormerod SJ. 2009. Climate change, river conservation and the adaptation challenge *Aquatic Conservation: Marine and Freshwater Ecosystems* **19**: 609-613.

Overland J, Wang M, Salo S. 2008. The recent Arctic warm period *Tellus A* **60**: 589-597.

Parker SM, Huryn AD. 2006. Food web structure and function in two arctic streams with contrasting disturbance regimes *Freshwater Biology* **51**: 1249-1263.

Parker SM, Huryn AD. 2011. Effects of natural disturbance on stream communities: a habitat template analysis of arctic headwater streams *Freshwater Biology* **56**: 1342-1357.

Payn RA, Webster JR, Mulholland PJ, Valett HM, Dodds WK. 2005. Estimation of stream nutrient uptake from nutrient addition experiments *Limnology and Oceanography: Methods* **3**: 174-182.

Pecher K. 1994. Hydrochemical analysis of spatial and temporal variations of solute composition in surface and subsurface waters of a high arctic catchment *Catena* **21**: 305-327.

Peckarsky BL. 1983. Biotic Interactions or Abiotic Limitations. A Model of Lotic Community Structure. In *Dynamics of Lotic Systems,* Fontaine III TD, Bartell SM (eds). Ann Arbor Science: Ann Arbor MI; 303-323.

Peterson BJ, Holmes RM, McClelland JW, Vorosmarty CJ, Lammers RB, Shiklomanov AI, Shiklomanov IA, Rahmstorf S. 2002. Increasing river discharge to the Arctic Ocean *Science* **298**: 2171-2173.

Peterson BJ, Wollheim WM, Mulholland PJ, Webster JR, Meyer JL, Tank JL, Martí E, Bowden WB, Valett HM, Hershey AE. 2001. Control of nitrogen export from watersheds by headwater streams *Science* **292**: 86-90.

Petrone KC, Jones JB, Hinzman LD, Boone RD. 2006. Seasonal export of carbon, nitrogen, and major solutes from Alaskan catchments with discontinuous permafrost *Journal of Geophysical Research* **111**: G02020.

Pfankuch DJ. 1975. Stream reach inventory and channel stability evaluation *US Department of Agriculture Forest Service, Region 1*: Missoula, MO, U.S.A.

Pohl S, Marsh P, Bonsal B. 2007. Modeling the impact of climate change on runoff and annual water balance of an Arctic headwater basin *Arctic* **60**: 173-186.

Poole GC, Berman CH. 2001. An Ecological Perspective on In-Stream Temperature: Natural Heat Dynamics and Mechanisms of Human-Caused Thermal Degradation *Environmental Management* **27**: 787-802.

Post E, Forchhammer MC, Bret-Harte MS, Callaghan TV, Christensen TR, Elberling B, Fox AD, Gilg O, Hik DS, Høye TT. 2009. Ecological dynamics across the Arctic associated with recent climate change *Science* **325**: 1355-1358.

Power G, Power M. 1995. Ecotones and fluvial regimes in Arctic lotic environments *Hydrobiologia* **303**: 111-124.

Prowse TD, Brown K. 2010. Hydro-ecological effects of changing Arctic river and lake ice covers: a review *Hydrology Research* **41**: 454.

Prowse TD, Wrona FJ, Reist JD, Gibson JJ, Hobbie JE, Lévesque LMJ, Vincent WF. 2006. Climate change effects on hydroecology of Arctic freshwater ecosystems *AMBIO: A Journal of the Human Environment* **35**: 347-358.

Putkonen J, Roe G. 2003. Rain-on-snow events impact soil temperatures and affect ungulate survival *Geophysical Research Letters* **30:** 1188.

Radić V, Hock R. 2011. Regionally differentiated contribution of mountain glaciers and ice caps to future sea-level rise *Nature Geoscience* **4**: 91-94.

Rahel FJ, Olden JD. 2008. Assessing the effects of climate change on aquatic invasive species *Conservation Biology* **22**: 521-533.

Rasband WS. 2012. *ImageJ* Available at: http://imagej.nih.gov/ij Accessed: 15/07/2012

Rasmussen JJ, Baattrup-Pedersen A, Riis T, Friberg N. 2011. Stream ecosystem properties and processes along a temperature gradient *Aquatic Ecology* **45**: 231-242.

Redfield AC. 1958. The biological control of chemical factors in the environment *American Scientist* **46**: 205-221.

Rennert KJ, Roe G, Putkonen J, Bitz CM. 2009. Soil thermal and ecological impacts of rain on snow events in the circumpolar Arctic *Journal of Climate* **22**: 2302-2315.

Richards J, Moore RD. 2011. Discharge dependence of stream albedo in a steep proglacial channel *Hydrological Processes* **25**: 4154-4158.

Rippin D, Willis I, Arnold N, Hodson A, Moore J, Kohler J, Björnsson H. 2003. Changes in geometry and subglacial drainage of Midre Lovénbreen, Svalbard, determined from digital elevation models *Earth Surface Processes and Landforms* **28**: 273-298.

Rossaro B. 1980. Description of some unknown larvae of *Diamesa* genus and corrections to previous descriptions (Diptera: Chironomidae) *Archiv für Hydrobiologie* **90**: 298-308.

Roth K, Boike J. 2001. Quantifying the thermal dynamics of a permafrost site near Ny-Ålesund, Svalbard *Water Resources Research* **37**: 2901-2914.

Rowland J, Jones C, Altmann G, Bryan R, Crosby B, Geernaert G, Hinzman L, Kane D, Lawrence D, Mancino A, Marsh P, Mcnamara J, Romanovsky V, Toniolo H, Travis B, Trochim E, Wilson C. 2010. Arctic landscapes in transition: Responses to thawing permafrost *Eos* **91.26:** 229-236.

Roy JW, Zaitlin B, Hayashi M, Watson SB. 2011. Influence of groundwater spring discharge on small-scale spatial variation of an alpine stream ecosystem *Ecohydrology* **4**: 661-670.

Runkel RL. 1998. One-Dimensional Transport with Inflow and Storage (OTIS): A Solute Transport Model for Streams and Rivers. U.S. Geological Survey Water-Resources Investigations Report: 98-4018.

Rutter N, Hodson A, Irvine-Fynn T, Solås MK. 2011. Hydrology and hydrochemistry of a deglaciating high-Arctic catchment, Svalbard *Journal of Hydrology* **410**: 39-50.

Säwström C, Laybourn-Parry J, Granéli W, Anesio A. 2007. Heterotrophic bacterial and viral dynamics in Arctic freshwaters: results from a field study and nutrient-temperature manipulation experiments *Polar Biology* **30**: 1407-1415.

Scanlon TM, Raffensperger JP, Hornberger GM. 2001. Modeling transport of dissolved silica in a forested headwater catchment: Implications for defining the hydrochemical response of observed flow pathways *Water Resources Research* **37**: 1071-1082.

Scordo EB, Moore RD. 2009. Transient storage processes in a steep headwater stream *Hydrological Processes* **23**: 2671-2685.

Scott D, Hood E, Nassry M. 2010. In-stream uptake and retention of C, N and P in a supraglacial stream *Annals of Glaciology* **51**: 80-86.

Screen JA, Simmonds I. 2010. The central role of diminishing sea ice in recent Arctic temperature amplification *Nature* **464**: 1334-1337.

Serreze M, Barrett A, Stroeve J, Kindig D, Holland M. 2009. The emergence of surface-based Arctic amplification *Cryosphere* **3**: 11-19.

Serreze M, Walsh J, Chapin FS, Osterkamp T, Dyurgerov M, Romanovsky V, Oechel W, Morison J, Zhang T, Barry R. 2000. Observational evidence of recent change in the northern high-latitude environment *Climatic Change* **46**: 159-207.

Serreze M, Francis JA. 2006. The Arctic Amplification Debate *Climatic Change* **76**: 241-264.

Sharp M, Ananicheva M, Arendt A, Hagen J, Hock R, Joseberger E, Moore R, Pfeffer W, Wolken G. 2011. Chapter 7: Mountain glaciers and ice caps in *Snow, Water, Ice and Permafrost in the Arctic (SWIPA)*. Arctic Monitoring and Assessment Programme (AMAP): Oslo.

Sharp M, Tranter M, Brown GH, Skidmore M. 1995. Rates of chemical denudation and CO_2 drawdown in a glacier-covered alpine catchment *Geology* **23**: 61-64.

Shaver G, Chapin F. 1980. Response to fertilization by various plant growth forms in an Alaskan tundra: nutrient accumulation and growth *Ecology* **61**: 662-675.

Shiklomanov I, Shiklomanov A. 2003. Climatic change and the dynamics of river runoff into the Arctic Ocean *Water Resources* **30**: 593-601.

Slavik K, Peterson B, Deegan L, Bowden W, Hershey A, Hobbie J. 2004. Long-term responses of the Kuparuk River ecosystem to phosphorus fertilization *Ecology* **85**: 939-954.

Smith LC, Pavelsky TM, Macdonald GM, Shiklomanov AI, Lammers RB. 2007. Rising minimum daily flows in northern Eurasian rivers: A growing influence of groundwater in the high-latitude hydrologic cycle *Journal of Geophysical Research* **112**: G04S47.

Soininen J. 2005. Assessing the current related heterogeneity and diversity patterns of benthic diatom communities in a turbid and a clear water river *Aquatic Ecology* **38:** 495-501.

St Jacques JM, Sauchyn DJ. 2009. Increasing winter baseflow and mean annual streamflow from possible permafrost thawing in the Northwest Territories, Canada *Geophysical Research Letters* **36**: L01401.

Steinman AD, Lamberti GA, Leavitt P. 2006. Biomass and pigments of benthic algae. In *Methods in stream ecology,* Hauer FR, Lamberti, GA (eds). Elsevier: USA; 357-379.

Stewart IT. 2009. Changes in snowpack and snowmelt runoff for key mountain regions *Hydrological Processes* **23**: 78-94.

Stibal M, Tranter M, Telling J, Benning LG. 2008. Speciation, phase association and potential bioavailability of phosphorus on a Svalbard glacier *Biogeochemistry* **90**: 1-13.

Stieglitz M. 2003. The role of snow cover in the warming of arctic permafrost *Geophysical Research Letters* **30**.

Stur E, Ekrem T. 2011. Exploring unknown life stages of Arctic Tanytarsini (Diptera: Chironomidae) with DNA barcoding *Zootaxa* **2743**: 27-39.

Sturt MM, Jansen MA, Harrison SS. 2011. Invertebrate grazing and riparian shade as controllers of nuisance algae in a eutrophic river *Freshwater Biology* **56:** 2580-2593.

Stutter M, Billett M. 2003. Biogeochemical controls on streamwater and soil solution chemistry in a High Arctic environment *Geoderma* **113**: 127-146.

Sueker JK, Ryan JN, Kendall C, Jarrett RD. 2000. Determination of hydrologic pathways during snowmelt for alpine/subalpine basins, Rocky Mountain National Park, Colorado *Water Resources Research* **36**: 63-75.

Tank J, Bernot M, Rosi-Marshall E. 2006. Nitrogen limitation and uptake In *Methods in stream ecology,* Hauer FR, Lamberti, GA (eds). Elsevier: USA; 213-238.

Tank JL, Dodds WK. 2003. Nutrient limitation of epilithic and epixylic biofilms in ten North American streams *Freshwater Biology* **48**: 1031-1049.

Tarnocai C, Canadell J, Schuur E, Kuhry P, Mazhitova G, Zimov S. 2009. Soil organic carbon pools in the northern circumpolar permafrost region *Global Biogeochemical Cycles* **23**: GB2023.

Telling J, Anesio AM, Tranter M, Irvine-Fynn T, Hodson A, Butler C, Wadham J. 2011. Nitrogen fixation on Arctic glaciers, Svalbard *Journal of Geophysical Research* **116**: G03039.

Theakstone WH. 2003. Oxygen isotopes in glacier-river water, Austre Okstindbreen, Okstindan, Norway *Journal of Glaciology* **49**: 282-298.

Tockner K, Pusch M, Borchardt D, Lorang MS. 2010. Multiple stressors in coupled river-floodplain ecosystems *Freshwater Biology* **55**: 135-151.

Townsend-Small A, Mcclelland JW, Max Holmes R, Peterson BJ. 2011. Seasonal and hydrologic drivers of dissolved organic matter and nutrients in the upper Kuparuk River, Alaskan Arctic *Biogeochemistry* **103**: 109-124.

Tranter M, Brown G, Sharp M. 1993. The Use of Sulfate as a Tracer for the Delayed Flow Component of Alpine Glacial Runoff *International Association of Hydrological Sciences Publication* **215:** 89-89.

Tranter M, Brown GH, Hodson AJ, Gurnell AM. 1996. Hydrochemistry as an indicator of subglacial drainage system structure: a comparison of alpine and sub-polar environments *Hydrological Processes* **10**: 541-556.

Tranter M, Sharp M, Lamb H, Brown G, Hubbard B, Willis I. 2002. Geochemical weathering at the bed of Haut Glacier d'Arolla, Switzerland—a new model *Hydrological Processes* **16**: 959-993.

Tye A, Heaton T. 2007. Chemical and isotopic characteristics of weathering and nitrogen release in non-glacial drainage waters on Arctic tundra *Geochimica et Cosmochimica Acta* **71**: 4188-4205.

Uehlinger U, Malard F, Ward J. 2003. Thermal patterns in the surface waters of a glacial river corridor (Val Roseg, Switzerland) *Freshwater Biology* **48**: 284-300.

Uehlinger U, Robinson C, Hieber M, Zah R. 2010. The physico-chemical habitat template for periphyton in alpine glacial streams under a changing climate *Hydrobiologia* **657**: 107-121.

Uehlinger U, Zah R, Burgi H. 1998. The Val Roseg Project: temporal and spatial patterns of benthic algae in an Alpine stream ecosystem influenced by glacier runoff *International Association of Hydrological Sciences Publication* **248:** 419-424.

van der Ploeg MJ, Haldorsen S, Leijnse A, Heim M. 2012. Subpermafrost groundwater systems: Dealing with virtual reality while having virtually no data *Journal of Hydrology* **475:** *42-52*.

van der Wal R, Hessen DO. 2009. Analogous aquatic and terrestrial food webs in the high Arctic: The structuring force of a harsh climate *Perspectives in Plant Ecology, Evolution and Systematics* **11**: 231-240.

van Vliet MTH, Ludwig F, Zwolsman JJG, Weedon GP, Kabat P. 2011. Global river temperatures and sensitivity to atmospheric warming and changes in river flow *Water Resources Research* **47:** W02544.

Vavrus SJ, Holland MM, Jahn A, Bailey DA, Blazey BA. 2012. Twenty-first-century Arctic climate change in CCSM4 *Journal of Climate* **25**: 2696-2710.

Vincent WF, Callaghan TV, Dahl-Jensen D, Johansson M, Kovacs KM, Michel C, Prowse T, Reist JD, Sharp M. 2011. Ecological implications of changes in the Arctic cryosphere *AMBIO: A Journal of the Human Environment* **40**: 87-99.

Wadham J, Cooper R, Tranter M, Hodgkins R. 2001. Enhancement of glacial solute fluxes in the proglacial zone of a polythermal glacier *Journal of Glaciology* **47**: 378-386.

Wadham J, Hodson A, Tranter M, Dowdeswell J. 1998. The hydrochemistry of meltwaters draining a polythermal-based, high Arctic glacier, south Svalbard: I. The ablation season *Hydrological Processes* **12**: 1825-1849.

Wagner BJ, Harvey JW. 1997. Experimental design for estimating parameters of rate-limited mass transfer: analysis of stream tracer studies *Water Resources Research* **33**: 1731-1741.

Walsh J, Anisimov O, Hagen J, Jakobsson T, Oerlemans J, Prowse T, Romanovsky V, Savelieva N, Serreze M, Shiklomanov I. 2005. Chapter 6: Cryosphere and hydrology in *Arctic Climate Impact Assessment*. Cambridge University Press: Cambridge.

Walsh J, Overland J, Groisman P, Rudolf B. 2011. Ongoing climate change in the Arctic *AMBIO: A Journal of the Human Environment* **40**: 6-16.

Walvoord MA, Striegl RG. 2007. Increased groundwater to stream discharge from permafrost thawing in the Yukon River basin: Potential impacts on lateral export of carbon and nitrogen *Geophysical Research Letters* **34:** L12402.

Ward J. 1994. Ecology of alpine streams *Freshwater Biology* **32**: 277-294.

Ward J, Malard F, Tockner K, Uehlinger U. 1999. Influence of ground water on surface water conditions in a glacial flood plain of the Swiss Alps *Hydrological Processes* **13**: 277-293.

Webb B, Nobilis F. 1994. Water temperature behaviour in the River Danube during the twentieth century *Hydrobiologia* **291**: 105-113.

Webb B, Zhang Y. 1997. Spatial and seasonal variability in the components of the river heat budget *Hydrological Processes* **11**: 79-101.

Webb B, Hannah DM, Moore RD, Brown LE, Nobilis F. 2008. Recent advances in stream and river temperature research *Hydrological Processes* **22**: 902-918.

Webster JR, Mulholland PJ, Tank JL, Valett HM, Dodds WK, Peterson BJ, Bowden WB, Dahm CN, Findlay S, Gregory SV. 2003. Factors affecting ammonium uptake in streams – an inter-biome perspective *Freshwater Biology* **48**: 1329-1352.

Westermann S, Lüers J, Langer M, Piel K, Boike J. 2009. The annual surface energy budget of a high-arctic permafrost site on Svalbard, Norway *The Cryosphere* **3**: 245-263.

White D, Hinzman L, Alessa L, Cassano J, Chambers M, Falkner K, Francis J, Gutowski WJ, Holland M, Holmes RM, Huntington H, Kane D, Kliskey A, Lee C, Mcclelland J, Peterson B, Rupp TS, Straneo F, Steele M, Woodgate R, Yang D, Yoshikawa K, Zhang T. 2007. The Arctic freshwater system: Changes and impacts *Journal of Geophysical Research* **112:** G04S54.

Whittaker RH. 1972. Evolution and measurement of species diversity *Taxon* **21:** 213-251.

Wiederholm T. 1983. The larvae of Orthocladiinae (Diptera: Chironomidae) of the Holarctic region: keys and diagnoses. Part 1 *Entomologica Scandinavica Supplement* **19**: 1-449.

Wipfli MS, Baxter CV. 2010. Linking ecosystems, food webs, and fish production: subsidies in salmonid watersheds *Fisheries* **35**: 373-387.

Wollheim WM, Peterson BJ, Deegan LA, Hobbie JE, Hooker B, Bowden WB, Edwardson KJ, Arscott DB, Hershey AE, Finlay J. 2001. Influence of stream size on ammonium and suspended particulate nitrogen processing *Limnology and Oceanography* **46**: 1-13.

Woo MK. 1983. Hydrology of a drainage basin in the Canadian High Arctic *Annals of the Association of American Geographers* **73**: 577-596.

Woo MK, Kane DL, Carey SK, Yang D. 2008. Progress in permafrost hydrology in the new millennium *Permafrost and Periglacial Processes* **19**: 237-254.

Woodward G, Perkins DM, Brown LE. 2010. Climate change and freshwater ecosystems: impacts across multiple levels of organization *Philosophical Transactions of the Royal Society B: Biological Sciences* **365**: 2093-2106.

Woodward GUY, Dybkjaer JB, Ólafsson JS, Gíslason GM, Hannesdóttir ER, Friberg N. 2009. Sentinel systems on the razor's edge: effects of warming on Arctic geothermal stream ecosystems *Global Change Biology* **16**: 1979-1991.

Wright N, Hayashi M, Quinton WL. 2009. Spatial and temporal variations in active layer thawing and their implication on runoff generation in peat-covered permafrost terrain *Water Resources Research* **45**: W05414.

Wrona FJ, Prowse TD, Reist JD, Hobbie JE, Lévesque LMJ, Vincent WF. 2006. Climate change effects on aquatic biota, ecosystem structure and function *AMBIO: A Journal of the Human Environment* **35**: 359-369.

Yang Y, Xiao H, Wei Y, Zhao L, Zou S, Yang Q, Yin Z. 2011. Hydrological processes in the different landscape zones of alpine cold regions in the wet season, combining isotopic and hydrochemical tracers *Hydrological Processes* **26**: 1457-1466.

Ye B, Yang D, Kane DL. 2003. Changes in Lena River streamflow hydrology: Human impacts versus natural variations *Water Resources Research* **39**: SWC8.

Ye S, Covino TP, Sivapalan M, Basu NB, Li HY, Wang SW. 2012. Dissolved nutrient retention dynamics in river networks: A modeling investigation of transient flows and scale effects *Water Resources Research* **48**: W00J17.

Yvon-Durocher G, Caffrey JM, Cescatti A, Dossena M, Del Giorgio P, Gasol JM, Montoya JM, Pumpanen J, Staehr PA, Trimmer M. 2012. Reconciling the temperature dependence of respiration across timescales and ecosystem types *Nature*.

Zarnetske JP, Gooseff MN, Brosten TR, Bradford JH, Mcnamara JP, Bowden WB. 2007. Transient storage as a function of geomorphology, discharge, and permafrost active layer conditions in Arctic tundra streams *Water Resources Research* **43:** W07410.

Zarnetske JP, Haggerty R, Wondzell SM, Bokil VA, González-Pinzón R. 2012. Coupled transport and reaction kinetics control the nitrate source-sink function of hyporheic zones *Water Resources Research* **48:** W11508.

Zemp M, Hoelzle M, Haeberli W. 2009. Six decades of glacier mass-balance observations: a review of the worldwide monitoring network *Annals of Glaciology* **50**: 101-111.

Zhang T, Barry RG, Knowles K, Heginbottom J, Brown J. 1999. Statistics and characteristics of permafrost and ground-ice distribution in the Northern Hemisphere *Polar Geography* **23**: 132-154.

Zuur AF, Ieno EN, Elphick CS. 2010. A protocol for data exploration to avoid common statistical problems *Methods in Ecology and Evolution* **1**: 3-14.

Appendix A1 – Published research

A copy of a *Hydrological Processes* journal manuscript published from research in Chapter 5 is presented.

Appendix A2 – Stage-discharge relationships

Stage-discharge relationships derived using salt dilution gauging (Hudson and Fraser, 2005) are presented. Site-specific rating curves were generated for sites V1, V2 and S2 in 2010 (Figure A2.1) and for sites V1, V2, N1, N2 and N3 in 2011 (Figure A2.1).

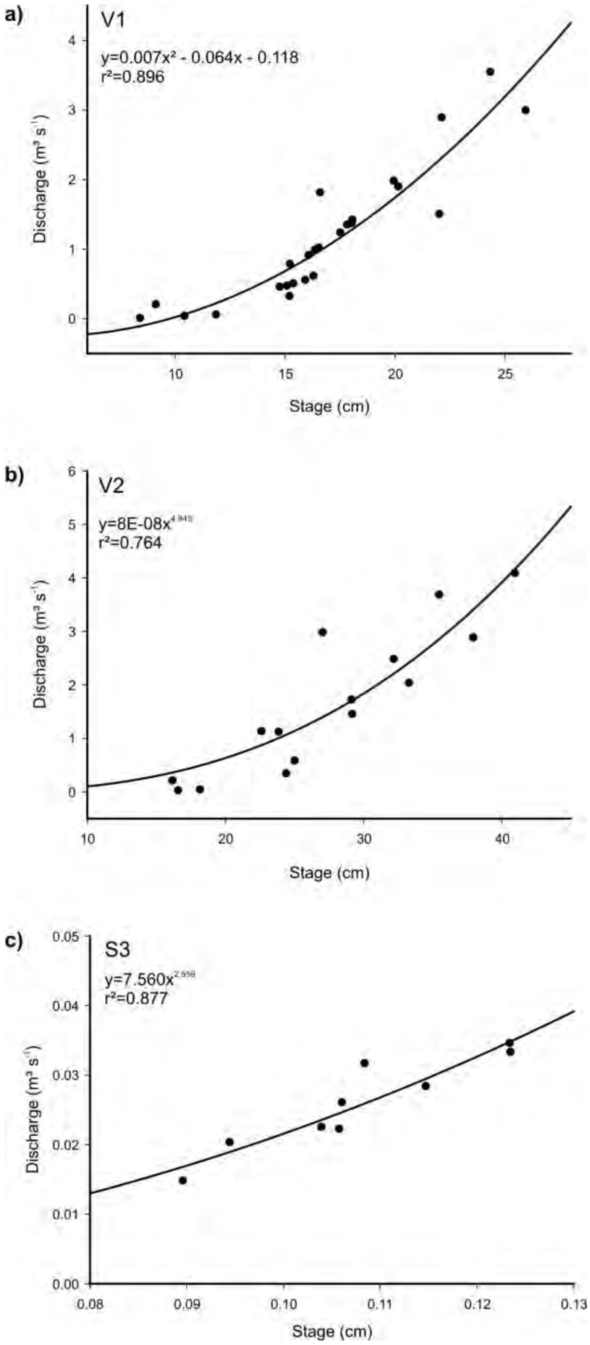


Figure A2.1: *Stage-discharge relationships for the 2010 field season at sites a) V1, b) V2 and c) S3*

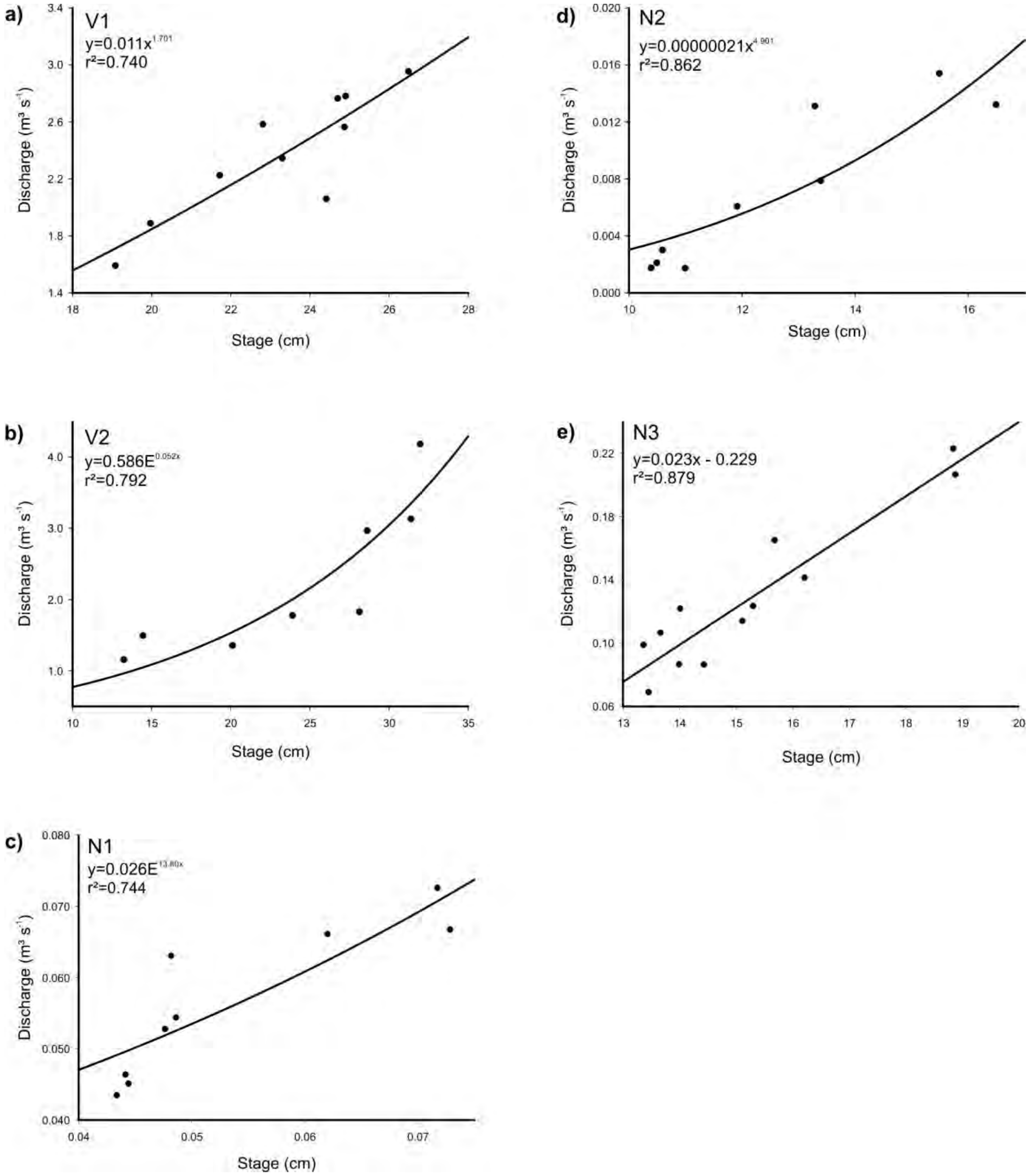


Figure A2.2: *Stage-discharge relationships for the 2011 field season at sites a) V1, b) V2, c) N1, d) N2 and e) N3*