THERMAL EVIDENCE FOR SURFACE AND SUBSURFACE WATER CONTRIBUTIONS TO BASEFLOW IN A HIGH ARCTIC RIVER

by

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A thesis submitted to the Department of Geography

In conformity with the requirements for

the degree of Master of Science

Queen’s University

Kingston, Ontario, Canada

(November, 2015)

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Abstract

Changes in river temperatures are caused by thermal energy exchanges at the interface between water and the atmosphere and between water, the streambed, and subsurface water. In permafrost regions, deeper active layer formation due to a warming climate can affect ground and channel bed thermal regimes and subsurface flow pathways. The main hypothesis is that within cold region landscapes, stream inflows into rivers provide relatively warm sources of water, while subsurface sources of water such as soil water contribute relatively cold water sources which alter the thermal and isotopic composition of the river, and as such, downstream temperature measurements can identify these sources in space and time.

In this study, river water temperature patterns were used as primary indicators of slope water exchanges along the West River at the Cape Bounty Arctic Watershed Observatory (CBAWO), Melville Island, Nunavut, Canada (75º N, 109º W). Water temperature data was collected through detailed longitudinal surveys along the river during the 2014 recession and baseflow periods to locate surface and subsurface lateral inflows. Limited water stable isotope sampling was also undertaken at fixed stations to determine possible mixing from different water sources.

Atmospheric factors and channel snow were found to be the main contributors to thermal variance in the river during the 2014 Summer season, with tributary inflow discharge also being a strong factor. The longitudinal temperature profiles indicate clear localized downstream changes in the thermal conditions of the river at multiple locations, and are interpreted to be indicative of subsurface and surface water exchange through inputs of cooler or warmer water. River temperature increased downstream and stable isotopic composition show progressive downstream enrichment in the two study reaches during the majority of the baseflow period,
which is indicative of a culmination of localized surface flow inputs along both reaches. Additionally, precipitation events may have increased local hillslope channel hydraulic gradients and therefore increased the inflow of older subsurface water from slopes to the river which subsequently changed the river’s isotopic composition over several days and resulted in downstream cooling.

These results demonstrate some key processes that influence the thermal regime of a High Arctic river and will contribute to a greater understanding of how surface, subsurface and other water exchanges influence stream hydrology, ecology and biogeochemistry.
Co-Authorship

This thesis and field research performed at the Cape Bounty Arctic Watershed Observatory (CBAWO) was supervised by Dr. Scott Lamoureux. Field measurements and sample collection was performed by the author with the help of his field assistant Nigel Bocking. All laboratory analysis and mapping work was also performed by the author while receiving assistance on the technical laboratory work aspects from Dan Lamhonwah. The thesis was written by the author, with editorial and scientific input and analysis expertise from Dr. Scott Lamoureux and Dr. Jan Franssen (Université de Montréal).
Acknowledgements

I would like to thank my supervisor, Dr. Scott Lamoureux for teaching me so much over the course of my studies at Queen's University. Without him I would not have been able travel to the High Arctic and have one of the most unique experiences in my life so far. Without him, this thesis would not have been possible, and I owe him much for helping to develop my curiosity and love of science. I have learned so much from you, and hope to be as knowledgeable as you are one day!

To everyone who I have met during my time in Kingston and at Queen's University, especially those whom I have shared a lab and a weather haven with, I sincerely thank you for your help, knowledge, patience and love, which has encouraged me to strive to be the best researcher I can be. I look forward to what all of you will accomplish over the years.

To my parents, Andre and Margot for their endless support and to my best friends, Collin and Kate whom gave me perspective and insight whenever I needed it, no words can explain how much I appreciate what you've done for me over the years. It is such an unbelievable feeling knowing that no matter what, I know I will always have your support and love. I look forward to more adventures together!

This research would not have been possible without the financial help and logistical support of ArcticNet, NSERC (Natural Sciences and Engineering Research Council), ADAPT, the Polar Continental Shelf Program, and Queen’s University.
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<td>ALD05 Creek</td>
</tr>
<tr>
<td>ALD</td>
<td>Active Layer Detachment</td>
</tr>
<tr>
<td>ASL</td>
<td>Above Sea Level</td>
</tr>
<tr>
<td>CBAWO</td>
<td>Cape Bounty Arctic Watershed Observatory</td>
</tr>
<tr>
<td>DEM</td>
<td>Digital Elevation Model</td>
</tr>
<tr>
<td>GMWL</td>
<td>Global Meteoric Water Line</td>
</tr>
<tr>
<td>GS</td>
<td>Goose Creek</td>
</tr>
<tr>
<td>IPCC</td>
<td>Intergovernmental Panel on Climate Change</td>
</tr>
<tr>
<td>MWI</td>
<td>Maximum Wetness Index</td>
</tr>
<tr>
<td>MX</td>
<td>Muskox Creek</td>
</tr>
<tr>
<td>PT</td>
<td>Ptarmigan Creek</td>
</tr>
<tr>
<td>TWI</td>
<td>Topographic Wetness Index</td>
</tr>
<tr>
<td>Q</td>
<td>Discharge</td>
</tr>
<tr>
<td>T</td>
<td>Temperature</td>
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<tr>
<td>WI</td>
<td>Wetness Index</td>
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<tr>
<td>WRGS</td>
<td>West River Gauging Station</td>
</tr>
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Chapter 1

Introduction

Mean annual surface air temperatures in the Arctic have increased at almost twice the global rate over recent decades and are predicted to increase by an additional 4-7°C over the next century (Frey and McClelland, 2009; AMAP, 2012; IPCC, 2014). Additionally, high latitude regions such as the High Arctic are likely to experience an increase in annual mean precipitation by the end of this century (IPCC, 2014). Therefore, warmer summers are expected to have deeper active layer formation (seasonal thaw layer) and permafrost degradation within Arctic regions. Hydrologically, one major consequence of a deeper active layer is the potential for increased exchange of water between the surface and subsurface, both in the soil and between rivers and other water bodies (Lawrence et al., 2008). Many of these exchanges will also potentially increase with additional summer rainfall events. As permafrost thaws, pathways are created within the active layer that direct precipitation, snowmelt and other water sources towards rivers and lakes. These pathways are controlled by various factors such as: soil composition, the presence of ice-rich layers, porosity, the presence of macropores, slope, aspect, vegetation and solar radiation. Once established, these pathways can alter water flow rates, and alter the fluxes of solutes and nutrients in the soil that were previously frozen as permafrost.

Subsurface water exchanges have been investigated in a limited way in High Arctic settings, particularly in light of changing the aforementioned permafrost conditions. Since changes in river temperatures are caused by thermal energy exchanges between water and the atmosphere, and between water and the streambed, fluctuations in river temperatures can be used to identify and localize these exchanges. This approach has been widely used in temperate
regions (Webb and Zhang, 1999; Westhoff et al., 2011). In cold region settings, these temperature variations are driven by injections of cooler water brought from snowmelt, precipitation and other subsurface sources (Blaen et al. 2012). In addition to the use of water temperature changes as a tracer of hydrological sources and pathways, the thermal dynamics of rivers can play an important role in altering physical and biogeochemical processes and aquatic ecosystem functions across a variety of spatial and temporal scales. Hence, there is a combined benefit to investigating the thermal regime of arctic rivers.

This research seeks to test the hypothesis that within cold region conditions and landscapes, surface inflows to river systems provide relatively warm sources of water, while subsurface sources of water such as soil water and/or hyporheic exchange in the channel bed provide colder sources of water relative to ambient river water temperatures. Inflows of these different water sources alter the thermal properties of the river, and as such, longitudinal temperature measurements can identify the influence of these sources in space and time. Additionally, stable isotope analysis of stream water provides a means to determine differing water contributions and can provide independent evidence of water source mixing derived from temperature measurements.

This research aims to provide a systematic analysis on such water exchanges that occur between surface water, subsurface water and river systems on a temporal and spatial scale through the use of thermal tracing. The goals of this study are to: (1) characterize seasonal patterns of temperature in response to seasonal hydrological change to obtain a baseline temperature of the river's thermal regime; (2) systematically examine longitudinal river temperature patterns; (3) identify water stable isotopic changes to determine downstream river
composition and source water changes; and (4) evaluate the role terrain plays in river
temperature dynamics in order to predict the likelihood of where such water exchanges occur.

In order to achieve these objectives, longitudinal river water temperature patterns were
used as a primary indicator of water exchanges along the 2.1 km lower reach of the West River
at the Cape Bounty Arctic Watershed Observatory (CBAWO), Melville Island, Nunavut, Canada
(75° N, 109° W). This catchment is characterized by shallow summer active layer development in
a continuous permafrost setting. This study attempts to determine some of the key processes that
influence the thermal regime of a high arctic river and will contribute to a greater understanding
of how water exchanges and terrain factors can influence stream hydrology, ecology and
biogeochemistry.
Chapter 2

Literature Review

2.1 Introduction

It has been well documented that climate change is being experienced more intensely in high latitude and high elevation settings (ACIA, 2005). If current trends continue, projected changes in these regions are expected to alter a number of key surface environments, and in particular, hydrological systems (ACIA, 2005; Frey and McClelland, 2009; AMAP, 2012; IPCC, 2014). As such, the Arctic environment will undergo significant changes, which will inevitably affect watershed hydrology and related landscape elements. Some of the projected changes include: a decrease in permafrost extent and thickness, deepening of the seasonal active layer, increases in summer precipitation, increased variability between surface and subsurface water flow pathways, and changes in thermal dynamics of river systems (Boulton et al., 1998; Zhang et al., 2006; Lawrence et al., 2008). This literature review synthesizes research related to watershed hydrology in Arctic settings, and how environmental and topographic factors influence the thermal regime of river systems and overall hydrological processes in the Canadian High Arctic.

2.2 Permafrost and active layer processes

Permafrost is usually defined as soil, rock or organic material that remains at, or below 0°C for at least two consecutive years (Woo, 2012). Geographically, the permafrost zones of Canada extend across a large proportion of the country's land area, with conditions varying from continuous to discontinuous patches (Figure 2.1) depending on latitude and elevation (Throop et al., 2012). Regional and local conditions affect the development of permafrost, with aspect, slope...
and soil type playing a role in controlling its spatial extent and depth (Woo, 2012). Regions in the High Arctic typically have a continuous distribution of permafrost. Zones of continuous permafrost are spatially represented where at least 90% of the area is covered by permafrost, while discontinuous or sporadic permafrost exists where the permafrost coverage varies between 10 and 90%. Permafrost is found to be a very strong control to the configuration of subsurface hydrological pathways, where discharge locations expand, as areas become freer of permafrost (Figure 2.1) (Walvood et al., 2012). Formation and changes in permafrost can greatly modify landscape and influence surface and subsurface hydrology, soil biogeochemistry, and ecosystems (Zhang et al., 2006).

![Image: The spatial distribution of permafrost in the northern hemisphere (Brown and Hannah, 2007).](image)

The active layer is the uppermost layer of the ground that typically undergoes annual thawing during the summer and freezing during the winter season. The freezing and thawing of the active layer can occur on a diurnal basis, which is typically the case in many temperate
regions, or on a seasonal basis in the high latitudes (French, 2007). Regardless of latitude, upper soil layers are usually most susceptible to heat fluctuations while lower soil layers are more thermally stable, due to progressive isolation from incoming solar radiation and atmospheric heat with increasing depth. Frozen soils can also be warmed by heat conduction and by heat transfer through water that is being infiltrated into soil such as rainfall and melting snow. A small amount of precipitation percolated into very cold soils can induce noticeable ground warming (Woo, 2012), which subsequently can lead to active layer deepening.

Specifically, the main factor consistently contributing active layer variations and ground temperature is air temperature, while precipitation and meltwater episodically affect active layer deepening (Oht, 2003). Depending on active layer depth, meltwater generated from snow cover can percolate through frozen soil, run off laterally or can be temporarily ponded above ground (Woo, 2012), contributing to potential heat conduction and heat transfer. Unfrozen water preserved in pockets within the active layer provides a conduit for rapid convective heat and water flow within soils (Boike et al., 1998). These active layer processes have major implications for the hydrological and thermal properties of river systems in the Canadian High Arctic.

2.3 Soil temperature

Soil temperatures in the High Arctic are more influenced by seasonality than lower latitude regions, particularly due to extended periods of daylight during the summer season. Soil temperatures are dependent on depth, because at the surface, temperature is controlled by the energy exchange with the atmosphere, and can change very rapidly. The energy balance at the earth’s surface, shown by Barry and Hare (1974) is as follows:

\[(1-\infty)K \downarrow + L \downarrow - L \uparrow = Q^* = Q_H + Q_E + Q_G\]  

(2.1)
where $\infty$ is surface albedo, $K_{\downarrow}$ is incoming short-wave radiation, $L$ is long-wave radiation, and the arrows indicate the upward or downward direction of the flux. $Q^*$ is net radiation, $Q_H$ is sensible heat flux, $Q_E$ is latent heat flux and $Q_G$ is heat flux into the ground.

These temperature fluctuations become progressively smaller with depth, due to the thermal diffusivity of the soil, and fluctuations depend on soil characteristics, soil moisture, climate, snow cover and vegetation (Figure 2.2) (Woo, 2012). The presence of snow typically mitigates ground to air temperature energy exchanges and limits surface temperatures to approximately 0°C during the snowmelt period. Specifically, an increase in snowfall in the early winter has a positive impact on the thermal regime of the soil, while late snowfall might have a negative impact because of its high albedo and consumption of latent heat during snowmelt (Oelke and Zhang, 2004).
At the surface above permafrost, heat fluctuations of soil temperatures can not only occur from above the ground, but also from the permafrost table upward (Woo, 1986; 2012). Under certain
conditions, there is the possibility that the seasonal freezing in each direction may not meet, which can leave an intervening thawed zone or *talik* between the two frozen zones (Woo 1986).

### 2.4 Thermal Regimes in Arctic Rivers

River temperatures are spatially and temporally variable according to a variety of conditions, with Arctic settings providing an additional set of factors that influence thermal dynamics in a hydrological system. Alpine systems are particularly sensitive because the timing, duration and magnitude of peak contributions from glacier melt, snowmelt, and precipitation are closely linked to climatic variability (Uehlinger et al. 2003; Brown et al. 2006, 2008; Dickson et al. 2012), which is also the case for Arctic systems. While there have been limited studies regarding the thermal dynamics of river systems in the High Arctic, in a more general context Caissie (2006) classified factors influencing river temperature into four main groups: 1) atmospheric conditions; (2) topography; (3) stream discharge; and (4) streambed (Figure 2.3). Of these four groups, atmospheric conditions are typically the most important factors and are mainly responsible for the majority of temperature exchange processes that occur at the surface of hydrological systems (Caissie, 2006; Webb 2008). However, Blaen et al. (2012) suggests that river temperatures are determined by not only hydroclimatological conditions, but also from various water sources.
Due to atmospheric conditions, the thermal regimes of rivers typically follow two major cycles at seasonal and diurnal scales. Over a season, broad thermal processes of river systems typically are divided into three consecutive stages. In the early season, river systems were found to be in the increasing temperature stage, during the mid-season the stable temperature stage, and the late season the decreasing temperature stage (Liu et al., 2005). These stages are mainly governed by changing atmospheric conditions throughout a given season (Caissie, 2006; Blaen et al., 2012). However, the increase and decrease in temperatures during the early and late stage are not constant across all river types. River size is also a factor towards overall downstream warming and cooling within these systems, with smaller rivers often having warming rates of 0.6°C km\(^{-1}\).
intermediate rivers $0.2^\circ$C km$^{-1}$, and larger rivers $0.09^\circ$C km$^{-1}$ (Zwieniecki and Newton, 1999; Torgerson et al., 2001; Caissie, 2006).

On a diurnal scale, river temperatures also have similar thermal stages compared to seasonal scales, but in a shorter timeframe. Over a 24 hour period, river temperatures typically increase during the morning, stabilize and peak during mid-day, and decrease during the late evening (Webband Zhang, 1999; Brown et al., 2005; Caissie 2006). As a result, water temperatures reach a daily minimum in the early morning, while the daily maximum temperature occurs in the late afternoon or early evening (Cassie, 2006). In high latitude regions such as the Canadian High Arctic, river temperatures have strong seasonality due to increases in atmospheric warming and solar warming during the summer season (Blaen et al., 2012).

While some rivers are thermally affected by groundwater, in Arctic settings, permafrost is broadly assumed to prevent deeper groundwater-surface water interactions, which limit the capacity of more thermally stable groundwater to limit or buffer river temperature variability (Brown et al., 2005; Blaen et al., 2012).

2.4.1 Location, geology, and topography affecting river temperature

Latitude and altitude also play a role in affecting river temperature dynamics. Research has shown that water temperature in Arctic regions is typically cooler and more stable compared to alpine rivers at lower latitudes, due to colder climate and reduced incoming solar radiation in high latitude regions (Blaen et al., 2012). While there is no tree cover in High Arctic settings, trees and mountains that shade river systems in other, lower latitude regions have been found to affect river temperature. Specifically, maximum river temperatures significantly decrease when the water surface is shaded, but minimum and mean temperatures are not substantially affected (Johnson, 2004; Brown and Hannah, 2008). However, the relative magnitude of minimum and
maximum temperatures are influenced by altitude. Mean water temperatures tend to decline with increasing altitude due to lower air temperatures and harsher climatic conditions that come with higher elevations (Webb and Walling, 1986; Brown and Hannah, 2008).

The geology of the landscape has also been found to be important when assessing river temperature variability. Streams and rivers flowing over bedrock were found to be responsive to solar radiation, which results in higher maximum and lower minimum temperatures, while water flowing over gravel-alluvial type substrate materials have dampened diurnal fluctuations in temperature due to longer hydraulic retention times (Johnson, 2004). Characteristics of the rock and sediment can also affect the degree of hyporheic exchange in a particular area, which in turn influences river temperatures. Fine-scale features such as grain size, shape, and composition of sediments and rocks determine most physical and chemical processes in the hyporheic zone (Zarnetske et al., 2007; Westhoff et al., 2011), in which sediment such as silt and clay are finer and tend to restrict or limit hyporheic exchange, while coarser sediment such as sand or gravel will allow a greater exchange (Boulton et al., 1998). Specifically, areas that exhibit hyporheic exchange have been found to have localized inputs of colder water (Zarnetske et al., 2008; Westhoff et al., 2011; Gariglio et al., 2013).

Topography also plays a large role in hydrologic connectivity within watersheds. For example, during periods of high wetness, hillslopes can be highly transmissive and help to contribute significant quantities of water to streams, rivers and areas adjacent to hillslopes (Jencso et al., 2009). Due to topographic controls such as slope, the expansion and contraction of lowland saturated zones is mostly determined by downslope redistribution of subsurface soil water (Stieglitz et al., 1997). Additionally, hillslopes help contribute wetness at its base, whereas soils on hillslopes tend to be progressively drier with elevation (Stieglitz et al., 1997; Jencso et
Regions underlain by permafrost also possess unusual drainage features named hillslope water tracks which also play a role in overall watershed connectivity (McNamara et al., 1999). Water tracks are essentially located in areas that have poorly defined depressions that overlay frozen ground and direct flow downslope, but do not have incised channels and are the dominant pathways for water removal from hillslopes in many Arctic Watersheds (McNamara et al., 1999). The presence of permafrost is essentially restricting the creation of incised channel networks by limiting erosion, and therefore the presence of water tracks occupy a flow regime that is the transition from areas within a landscape that are less conducive to flow to clearly incised channels that have universal wetness attractors (McNamara et al., 1999). Therefore, areas which do not have strong hillslope relief which direct flow similar to Alpine systems can still have localized hydrological connectivity through topographic features like water tracks (Figure 2.4).
2.4.2 River Heat Exchanges

Similar to soil surface temperatures, river temperatures can be highly variable on a daily and seasonal timescale due to atmospheric conditions and hydrological factors. Gu and Li (2002) proposed a generalized river temperature model, which can estimate mean river temperatures on hourly, daily and seasonal timescales across various river distances and reaches. This temperature equation is expressed as:

\[ T = T_e + (T_e - T_o) \frac{Q}{\kappa W} \left[ e^{-\left(\frac{\kappa W}{Q}\right)} - 1 \right] \]  \hspace{1cm} (2.2)
where $\alpha = KL/pc_p$ where $L$ is river reach length in meters, $p$ is water density, $K$ is the heat exchange coefficient (Figure 2.5), $c_p$ is heat capacity of water in kcal/kg°C, $Q$ is flow rate in $m^3/s$, $W$ is the river surface width, $T_e$ is the equilibrium temperature approximated by $T_e = T_d + \frac{R}{K}$, $R$ being solar radiation in W/m$^2$ and $T_d$ is the dew point temperature in °C, $T_o$ is upstream river temperatures in °C and $T$ is the daily mean river temperature in °C.

<table>
<thead>
<tr>
<th>Parameter or coefficient</th>
<th>Expression</th>
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<tbody>
<tr>
<td>Heat exchange coefficient, $K$</td>
<td>$K = 4.5 + 0.05T + \beta f(U_w) + 0.47f(U_w)$</td>
</tr>
<tr>
<td>Wind function, $f(U_w)$</td>
<td>$f(U_w) = 9.2 + 0.46U_w^2$</td>
</tr>
<tr>
<td>Coefficient, $\beta$</td>
<td>$\beta = 0.35 + 0.015T_v + 0.0012T_v^2$</td>
</tr>
<tr>
<td>Average temperature, $T_v$</td>
<td>$T_v = (T + T_d)/2$</td>
</tr>
<tr>
<td>Dew point temperature, $T_d$</td>
<td>$T_d = 237.3{T_{a^<em>} + \ln(r_h)/[17.27 - \ln(r_h) - T_{a^</em>}]}$</td>
</tr>
</tbody>
</table>

Where $T_{a^*} = 17.27T_d/(237.3 + T_d)$.

Note: $U_w =$ wind speed (m/s), $T_a =$ air temperature (°C), $T =$ unknown river temperature (°C), and $r_h =$ relative humidity.

**Figure 2.5: List of empirical formulas for computing surface heat exchange coefficient $K$, (Gu and Li, 2002).**

Webb and Zhang (1999) also proposed a conceptual heat budget to study reach temperature, which consists of:

$$Q_n = \pm Q_r \pm Q_e \pm Q_h \pm Q_{hb} + Q_{fc} + Q_p + Q_g$$

Where $Q_n$ is the total net heat exchange, $Q_r$ is the heat flux due to net radiation, $Q_e$ is the heat flux due to evaporation and condensation, $Q_h$ is the heat flux due to sensible heat transfer via convection and conduction between air and water, $Q_{hb}$ is the heat flux due to bed conduction, $Q_{fc}$ is the heat flux due to friction, $Q_p$ is the heat flux due to precipitation, and $Q_g$ is the heat flux due to groundwater. While these equations are useful to calculate river temperatures across reaches, it is important to note that none of these equations take into consideration incoming tributary
inflows, which considerably affect river temperatures (Cassie, 2006; Blaen 2012; Dingman 2012; Woo 2012).

Overall, heat transfers within river systems are complex, combining radiation, conduction, convection and advection. These types of energy exchanges can add or remove heat from a river system. Specifically, heat inputs can be caused by solar radiation, condensation, friction at the channel beds/banks, and chemical/biological processes (Brown et al., 2005; Caissie, 2006; Webb et al., 2008). Losses may include reflection of solar radiation, emission of longwave solar radiation, evaporation, and inputs from cooler sources such as subsurface inflows and channel snow (Blaen et al., 2012). Sensible heat and water column-bed energy transfers may also cause gains or losses. Additionally, other ways in which heat or energy can be lost/gained is through advection by incoming and outgoing stream discharge, evaporation, groundwater upwelling and downwelling, tributary inflows and precipitation events (Brown et al., 2005).

2.4.3 Surface inflows

Streamflow in the Arctic is typically sustained by several water sources consisting of spring snowmelt, rainfall, and the melting of semi-permanent snowbanks and glaciers. During the snowmelt period, diurnal cycles of runoff reflect the influence of solar energy timing and the resultant snowmelt input. Typically, seasonal snowmelt represents the main supply of water in the High Arctic. The shallow active layer in permafrost terrain usually cannot store the water released from snowmelt and other sources which results in a period of overland surface flow. Surface inflows can be separated into two categories called Hortonian overland flow and saturation overland flow. Hortonian overland flow occurs when the water supply exceeds the rate of infiltration resulting in excess runoff on the land surface, while saturation overland flow is a
mixture of precipitation and return flow, which occurs when the water table rises to the soil surface to generate runoff on the land surface (Woo, 2012). Surface inflows in the High Arctic typically are more common early during the melting season, due to frozen soils limiting infiltration and available water supply. The direction in which surface inflows follow is dependent on local topography and follows slope concavities, water tracks and defined channels that have cut into the slope of the landscape (Woo, 2012; Dingman, 2015). Similar to river systems, surface inflow temperatures are strongly affected by atmospheric conditions, although due to smaller discharge levels, are typically more sensitive to thermal inputs (Caissie, 2006). As such, surface inflows generally contribute warmer sources of water to river systems due to atmospheric energy gains (Gu and Li, 2002).

2.4.4 Subsurface inflows

Subsurface inflows typically occur in the saturated zone between the water and frost tables (Woo, 2012). While surface inflows cease as the water table deepens during the dry summer period, subsurface flow can continue beneath the surface. On Cornwallis Island, Nunavut, for example, surface flow is typically an order of magnitude or more in terms of flow compared to corresponding subsurface flow, particularly during spring snowmelt (Figure 2.4) (Nuttal and Callaghan, 2000).
Figure 2.6: Relative magnitude of surface and subsurface inflows on a slope in a yearly timescale (Nuttal and Callaghan, 2000).

These specific types of inflows have been shown to exhibit dampened diurnal fluctuations compared to surface inflows due to mitigated input from atmospheric conditions (Johnson, 2004; Caissie, 2006), and therefore, Blaen et al. (2012) suggest that subsurface inflows can provide inputs of colder water into hydrological systems. Similar to surface inflows, subsurface inflows tend to rise and fall sharply in terms of discharge, responding quickly to contributions of meltwater and precipitation, however, subsurface flow has a much longer duration than surface flow (Nuttal and Callaghan, 2000). While subsurface inflows also are affected by topography such as slope, unevenness of the frost table can deflect or block subsurface flow which would not entirely relate to surface gradients such as channelization or slope (Woo, 2012). In addition to cold water inputs, subsurface flow contributions to river systems can also contribute a substantial influx of nutrients, carbon and nitrogen (Alexander and Caissie, 2003).
2.4.5 Hyporheic exchanges

Hyporheic exchanges have been investigated in a limited way in Arctic settings, particularly in context of changing permafrost conditions. These exchanges occur in what is called the hyporheic zone, which is delineated by the volume of saturated sediment that surround a river system. This area is a hydrological transition between surface and subsurface waters within the riparian zone (Boulton et al., 1998). Specifically, hyporheic exchange occurs through circulation cells that move river water in downwelling and upwelling motions (Figure 2.5), with further interactions with groundwater (where present) located within the hyporheic zone (Boulton et al., 1998; Zarnetske et al., 2007; Westhoff et al., 2011). This upwelling and downwelling circulation tends to take place through a sequence of pools and riffles (Figure 2.5). Such a sequence is characterized through alternating areas of shallow riffle water and deeper pools in a river system (Storey et al., 2003). Previously, hyporheic flows were considered to have a very minor impact on stream temperature dynamics, however recent literature suggests that the hyporheic zone plays an important role in influencing stream temperature in alluvial reaches (Johnson, 2004; Caissie, 2006). Water mixing in the hyporheic zone can dampen stream water temperature variations at hourly and seasonal timescales (Westhoff et al., 2011).
In arctic river systems, hyporheic processes tend to be controlled by channel morphology influenced by ice and permafrost, where hyporheic exchange occurs when streambeds thaw seasonally (Zarnetske et al., 2008). However, hyporheic exchange is not only controlled by channel morphology, but is also influenced by discharge, slope, elevation of the channel stage, and riverbed permeability (Zarnetske et al., 2007; Westhoff et al., 2011). For example, as river flow increases the channel bed exerts less influence on the water-surface profile which causes subsurface-to-surface water exchanges to become more uniform (Tonina and Buffington, 2009), which is visually illustrated in Figure 2.6.
Figure 2.8: Hyporheic path lines in a pool-riffle channel with a) low discharge and b) high discharge (Tonina and Buffington, 2009).
The rate of mixing and the composition of sediment in the hyporheic zone can influence temperature, pH, electrical conductivity, and dissolved element concentrations in which these changes can be measured and used to determine the extent and influence of hyporheic interactions (Boulton et al., 1998; Tonina and Buffington, 2009). The effects of hyporheic exchanges have been known to have an impact on fish and macroinvertebrates (Boulton et al., 1998; Tonina and Buffington, 2009). Hyporheic processes have been recorded to be even more important within arctic settings due to the movement of nutrients and dissolved elements since nutrients tend to be much more limited in Arctic river systems (Zarnetske et al., 2007). This is because as water moves through sediment, there is a high opportunity for biogeochemical processing of dissolved material (Edwardson et al., 2003).

2.5 Precipitation events and river temperature

Since precipitation has the potential to rejuvenate surface inflows and affect the amount of volume of discharge in river systems, contribution of precipitation towards river temperatures should be considered. Brown and Hannah (2007) suggested that stream temperature responses to precipitation results from advective energy inputs which are primarily from surface and near surface hillslope pathways and by groundwater, rather than heat fluctuations from direct precipitation. Surface inflows that are rejuvenated by precipitation go through the same atmospheric conditions as meltwater flowing through streams, which results in localized inputs of warm water when solar radiation is present (Uehlinger et al., 2003; Caissie, 2006; Webb et al., 2008). Similarly to some alpine settings, precipitation can cause pre-event water within the
ground to be displaced or hydraulically pistoned within shallow soil (Brown and Hannah, 2007), which can cause localized injections of cooler water within a river or stream (Blaen et al, 2012).

### 2.6 Water stable isotope composition

In most low temperature environments, stable isotopes such as δD and δ¹⁸O behave conservatively, and interactions with other organic and geologic materials will have a negligible effect on isotopic composition. The main hydrological processes that affect changes in isotope composition in water are: (1) phase changes such as snowmelt, evaporation or condensation; and (2) mixing of water between the surface and the subsurface (Friedman et al., 1984). Stable isotopes in water can be used to partition source water contributions, especially during periods of high runoff since rainfall, snowmelt and other source water compositions are generally isotopically different (Friedman et al. 1984; Obradovic and Sklash, 1986). Isotope hydrograph separation is a technique used to partition these source water contributions, namely event water such as rain or snowmelt and pre-event water such as groundwater, and soil water (Taylor et al., 2002). Essentially, isotope hydrograph separation is based on the assumption that the isotopic composition of water in a watershed before an event (such as soilwater) and during an event (rainfall) is known at a specific point in space and time, and that river water is a mixture of event and pre-event composition (Taylor et al., 2002). Therefore, identifying various isotopic compositions within a watershed can help identify where source water mixing within a river is coming from, and how much influence these source water contributions have when mixing into a river or a stream.

The Global Meteoric Water Line (GMWL) was developed as an overall baseline for stable isotopic compositions (Friedman et al. 1984; Obradovic and Sklash, 1986), although Local
Meteoric Water Lines (LMWL) can have slightly different isotopic patterns than the global line to better represent whether regional isotopic compositions are more enriched or depleted in comparison to global averages. Additionally, these meteoric lines also help to determine whether various source water contributions, whether event or pre-event are more enriched or depleted in comparison to the local mean composition.

For example, Figure 2.8 demonstrates how various hydrological processes affect $\delta^D$ and $\delta^{18}O$ in relation to global and local meteoric water lines. Cooler, high latitude areas located in the High Arctic would have more depleted isotope compositions, relative to warm, low latitude areas (Clark and Fritz, 1997). Similar to latitudinal variations, there is seasonal variation of isotopic composition due to seasonal variation in air temperature, with summer being more enriched and winter having a more depleted state. These differences in isotopic composition can help to determine whether various water sources are from pre-event contributions (previous winter soilwater/ice) or event contributions (summer season precipitation) (Clark and Fritz, 1997; Taylor et al., 2002).
Figure 2.9: Summary diagram of how hydrological processes can affect isotope composition (Adapted from: Clark and Fritz, 1997).

However, isotopic hydrograph separations do not completely explain the flow paths that event and pre-event water sources follow to reach the river channel (Kendall and McDonnell, 2012), which reinforces that other analyses such as thermal evidence and topographic analysis should take place in order to more fully understand the nature of source water contributions in space and time.
2.7 River Temperature and ecosystem health

Thermal dynamics of rivers can play an important role in altering physical and biogeochemical processes and aquatic ecosystem function across a variety of spatial and temporal scales. There are many biological factors and conditions, in addition to stream productivity, that are strongly linked to stream water temperature. Specifically, river thermal shifts can affect species abundance and diversity in aquatic benthic communities (Brown and Hannah, 2007; Blaen et al., 2012). Fish and other aquatic biota have specific temperature preferences which can determine their distribution within variable river systems. Regions of localized subsurface discharge in streams have been known to provide thermal refuge for fish spawning and nursery habitats (Alexander and Caissie, 2003). Caissie (2006) notes that water temperature is very important for salmon growth, for the timing of fish movement, emergence of fish, and spawning. Biological productivity within streams are also significantly related to water temperature.

In general, approximately every ~10°C difference in water temperature results in a doubling of biological productivity (Brown and Krygier, 1967). While temperature increases might increase biological activity, temperatures between 23 to 25°C affect trout mortality, salmon are affected at slightly higher temperatures of 27 to 28°C (Caissie, 2006), and Northern Pike at 25°C (Reist et al., 2006). Unlike temperate and alpine regions, many Arctic settings do not have forest canopy to shade river systems, and as such, fish have little cover from solar radiation when migrating in river systems. As mean temperatures will likely increase due to climate change (Frey and McClelland, 2009; AMAP, 2012; IPCC, 2014), the role of colder inputs of subsurface waters (Carrivick et al., 2012; Blaen et al., 2012) will serve to a play strong
role for thermal refuges during spawning, feeding and migration to limit fish mortality based on
the previously mentioned thresholds. Although many projected effects of climate change on
arctic freshwater fishes seem to be generally positive, such as increased growth due to increasing
temperatures, favourable changes in other categories such as nutrient loading and productivity of
lower trophic organisms must also occur (Reist et al., 2006).

2.8 Conclusion

Projected climate change inducing global warming trends will cause significant changes
to arctic landscape and hydrological systems (ACIA, 2005; Frey and McClelland, 2009; AMAP,
2012; IPCC, 2014). Additionally, increases in annual precipitation in the High Arctic will lead to
rejuvenation of stream inflows during baseflow and will also affect the dynamics of the active
layer and distribution of permafrost. These changes in permafrost will modify landscapes and
influence surface and subsurface hydrology, soil biogeochemistry, and local ecosystems (Zhang
et al., 2006). Thermal dynamics within river systems are affected by a multitude of factors, with
meteorological and stream discharge, such as surface and subsurface flow being the major
contributors to thermal energy exchanges. Topography can also influence stream temperature,
but mostly serves to limit or redirect the location of thermal energy exchanges between water
sources. The timing and magnitude of water temperature fluctuations within river systems in the
High Arctic will challenge the resiliency of local wildlife such as fish spawning and
microorganisms within river systems. Effects of climate change could significantly modify the
distribution of aquatic organisms as water temperature in some systems is already reaching the
lethal limit for fish (Eaton et al., 2005).
Chapter 3

Thermal Evidence for Surface and Subsurface Water Contributions to Baseflow in a High Arctic River

3.1 Abstract

Spatial and temporal river water and bed temperature patterns were used as thermal evidence for slope-river water exchanges along the length of the West River at the Cape Bounty Arctic Watershed Observatory (CBAWO), Melville Island, Nunavut, Canada (74°55' N, 109°35' W). Water temperature data was collected through fine-scale 1 second resolution longitudinal surveys along the river for the two month snowmelt recession and baseflow period during the 2014 summer season to locate lateral inflows from the surface and subsurface. Water stable isotope sampling was also undertaken to determine the extent of surface water mixing from different sources.

Atmospheric factors were found to be the main contributors to seasonal thermal variance in the river, with inflow discharge also being a strong contributor. The longitudinal temperature profiles indicate notable changes to the thermal state of the river at multiple locations, some of which are interpreted to be indicative of subsurface to surface water exchange through inputs of relatively cold or warm water. Broadly, temperature increased downstream and stable isotopes indicated progressive enrichment in two study reaches during much of the baseflow period. Further, the late season depletion of river stable isotopic composition suggests that rain events, likely increased local hillslope channel hydraulic gradients which increased the flux of depleted subsurface water to the river over several days and resulted in downstream cooling.
These results demonstrate some key processes that influence the thermal regime of a High Arctic river and will contribute to a greater understanding of surface, subsurface and other water exchanges that influence stream hydrology, ecology and biogeochemistry.

3.2 Introduction

Current scientific consensus indicates that global temperatures will increase, with most warming occurring in high latitude regions (ACIA, 2005; Frey and McClelland, 2009; IPCC, 2014). These higher temperatures will, on average, lengthen the summer thaw season, which will contribute to seasonal active layer deepening in permafrost regions. Additionally, rainfall is projected to increase in terms of frequency and magnitude in high latitude regions (AMAP, 2012; IPCC, 2014). These changes have implications for watersheds in the Canadian High Arctic as active layer deepening can affect channel thermal regimes and subsurface flow pathways on the land. These pathways within the active layer can direct precipitation, snowmelt and other water sources towards channels and other surface water bodies (Woo and Steer, 1983; Woo, 2012). Despite their importance in the thermal and biogeochemical conditions in rivers in temperate settings, the processes that control exchanges between surface, subsurface and hyporheic waters have been investigated in a much more limited way in Arctic settings (Zarnetske et al., 2007; Westhoff et al., 2011), particularly during recent developments concerning Climate Change, characterized by projections of warmer summer temperatures and changing permafrost conditions. Since changes in river temperatures are caused by thermal energy exchanges between the atmosphere, water, streambed, and water inflows, fluctuations in river temperatures can be used to identify and localize these latter exchanges (Carrivick et al.,
In cold region settings, these temperature variations are likely to be caused by inflows of cooler water brought from snowmelt, precipitation, and particularly through subsurface soil flow and other shallow groundwater sources if present (Blaen et al. 2012). However, to date this approach has not been undertaken in an Arctic setting.

This research aims to test the hypothesis that subsurface water exchanges in an Arctic river can be identified on the basis of longitudinal water temperature changes. The objectives of this study were to: (1) characterize seasonal patterns of temperature to obtain a baseline temperature of the river's thermal regime; (2) examine longitudinal river temperature dynamics in order to identify the spatial and temporal departures consistent with water exchanges; (3) characterize different source waters, if possible, through stable isotope analysis (δD and δ\(^{18}\)O); and (4) evaluate terrain controls over observed patterns of water inputs determined from the second objective.

### 3.3 Study Site

Research was carried out at the Cape Bounty Arctic Watershed Observatory (CBAWO), Melville Island, Nunavut, Canada (74°55' N, 109°35' W) (Figure 3.1). CBAWO was established in 2003 in order to investigate climate change influences on hydrological processes, permafrost, soils, vegetation, greenhouse gas emissions, and contaminant cycling, particularly in High Arctic rivers and lakes. It is composed of similar paired watersheds (East and West Rivers, unofficial names, respectively) which flow into similar downstream lakes. The main focus of this study is the lower reaches of the West River, a 2.1 km-long channel that drains an 8.0 km\(^2\) watershed. This river has a moderate gradient profile, with a mean slope of 0.015 that decreases slightly downstream towards West Lake (Veillette, 2011). The watershed is underlain by Devonian
sandstone and siltstone bedrock, and draped with Quaternary marine and glacial sediments (Hodgson et al., 1984; Lamoureux and Lafrenière, 2009). Both the upper and lower reaches of this river system is characterized by many series of pools and riffles. Topography is characterized by dissected uplands that result in rolling terrain with gentle slopes. Drainage within these slopes have shallow channels or diffuse flow in water tracks over saturated soils (Lamoureux and Lafrenière, 2009).

The climate of the area is a polar semi-desert which typically has long cold winters and a short melt summer season, with mean July temperatures from 2003 to 2012 averaging 6.6°C, and mean summer precipitation during that same time period averaging 26.3 mm. (Favaro and Lamoureux, 2014). Most of the annual precipitation occurs in the form of snow. Typically, stream flow begins in mid-June, and discharge rapidly reaches a well-defined peak within a few days. Rapid depletion of snow subsequently leads to flow recession and baseflow with sporadic rainfall runoff events, and continues until autumn freeze up in late August or early September (Lewis et al., 2012).

Continuous, thick permafrost underlies the landscape in this region with a seasonal active (thawed) layer that can reach 0.5-1.0 m by late summer, depending on the temperature conditions for the given year (Rudy et al., 2013). Soils are characterized by thin cryic regosols, which tend to be saturated during and shortly after the snowmelt period, particularly in areas with low slopes and where topography focuses drainage (Lewis et al., 2012). The landscape is characterized by tundra vegetation that varies in composition and cover primarily based on soil water availability.
Figure 3.1. A portion of the Cape Bounty Arctic Watershed Observatory (CBAWO) study site showing a 2.1 km reach of the West River. Stations referred to in the text are indicated: meteorological station (MainMet), and river sampling stations (Station 4-11, Alpha, Beta). Tributary stream catchments mentioned in the text are also indicated. Contour intervals are in meters above sea level and are sourced from a DEM. The red circle within the inset map shows the location of CBAWO in the Canadian High Arctic.
3.4 Methods

3.4.1 Meteorology and hydrology

Meteorological data was collected from the station MainMet that was established in 2003 (Figure 3.1). Air temperatures were measured at 1.5 m above ground with a shielded Onset temperature-relative humidity sensor (± 0.2°C temperature, 5% RH) and rainfall was measured with an Onset tipping bucket gauge (0.2 mm tip). Both were logged hourly with an Onset U30 logger. Additional soil temperatures were measured at the Lower Goose station (Figure 3.1) at depths of 5, 15, 40 and 65 cm with Decagon 5TE sensors and a Em50 data logger at 3 hour intervals (0.1°C accuracy).

Hydrometric data was collected at the West River gauging station (WRGS) during June and July 2014 with missing data on June 26th due to maintenance. Stage was measured with a Unidata capacitive water depth probe logged with an Onset H22 logger at 10 minute intervals. The river was manually rated throughout the season with a Swoffer 2100 current velocity meter (1% accuracy) to measure cross-sectional velocity, and discharge (Q) was calculated by the velocity-area method using the following:

\[ Q = \sum \left( \Delta x \ast \left( \frac{h_1+h_2}{2} \right) \ast \left( \frac{v_1+v_2}{2} \right) \right) \]  \hspace{1cm} (3.1)

where \( h \) is the water depth (m), \( v \) is the flow velocity (m/s), and \( \Delta x \) is the horizontal interval between measurements. An exponential discharge rating best fit curve was calculated (\( r^2=0.90, \) \( n=42 \)):

\[ Q = 16.77h^2 - 2.533h \]  \hspace{1cm} (3.2)

where \( h \) is the recorded stage (m) and subsequently applied to the stage depth measurements collected at the gauging station. This rating curve used velocities from the West River ranging
from 0.02 m/s to 1.52 m/s. Tributaries flowing into the West River were also gauged for the summer season using calibrated cutthroat flumes and rated with calibrated standard discharge curves (Lamoureux et al., 2014).

3.4.2 River temperature

Multiple longitudinal water temperature tow surveys were undertaken after the main snowmelt period to measure variations in temperature indicative of different water inflows. Conditions during snowmelt were considered to be dominated by snowmelt runoff in the river and flow and channel access (due to slush and deep channel snow) was not suitable for consistent tows during this time.

Weekly surveys were carried out in the upstream direction on foot using a highly sensitive temperature sensor and logger unit (Richard Brancker Research, Ottawa, Canada) RBR model 1060F (90 ms response time, calibrated accuracy of <0.001°C) to measure longitudinal channel temperature at 1 second intervals. This logger was towed in a miniature float configuration handled by a 2 m fishing pole and line to separate the operator and to avoid influencing measurement conditions. The float was also equipped with a Garmin 76CSx GPS (typical accuracy of 3m) to position the temperature measurements at 3 second intervals (approximately 3-5 m distance). The location of 1 second temperature measurements were linearly extrapolated from the 3 second GPS locations and represented a mean measurement interval of 1-2m. An Onset UA-004 3-axis tilt logger was also secured to the float to time stamp when the float and temperature sensor were forced out of the water due to bed irregularities and residual ice and snow. These tow surveys were conducted in the morning and in the afternoon on July 6, 10 and 19 at 9-11 AM and 4-6 PM, corresponding to approximate daily low and high
flow conditions, respectively, as well as during a rainfall runoff event on July 16. Tows were not possible after July 19 due to low water levels.

In order to identify subtle longitudinal changes in temperature on a longitudinal scale, $\Delta T_i$ was calculated between each subsequent temperature measurement interval along the tow surveys by:

$$\Delta T_i = T_1 - T_2 \ldots T_n - T_{n+1}$$

3.3

where $T_1$ is the first upstream temperature value and $T_2$ is the immediate downstream temperature measurement. In addition, the cumulative departure for $\Delta T_i$ was calculated downstream (by summing adjacent $\Delta T_i$ values) to more easily define and show progression of losses and gains in temperature between reaches.

Water temperature was also measured at fixed stations 4-11 (Figure 3.1) with paired Onset UA-002 pendant loggers (accuracy ± 0.5°C) inserted into the streambed at 5 cm depth, and 5 cm above the bed in the water to determine differentials related to changing water sources. Loggers were covered with white electrical tape to minimize solar heating effects.

3.4.4 Stable isotope analysis

Water samples were collected for stable water isotope analysis ($\delta^D$ and $\delta^{18}O$) to distinguish the mixing of different water sources in the river. Water samples were collected at fixed locations during longitudinal tow surveys on July 6th, 10th, 16th and 19th during both the morning and the afternoon (Figure 3.1). Samples were also collected on July 13th, 22nd and 24th without corresponding tow surveys (due to low discharge) and also from the outlet of named tributaries throughout the season (Figure 3.1). Samples were filtered through 0.2 µm Osmonics filters and retained in 25ml scintillation vials without headspace for subsequent analysis.
Isotope analysis was conducted with a Los Gatos Research (LGR) analyzer at Queen’s University that utilizes Off-Axis Integrated Cavity Output Spectroscopy (OA-ICOS). The uncertainty ranges of this instrument are ± 0.2‰ for δ¹⁸O and ±0.8‰ for δD, respectively. Each sample is analyzed by the LGR with six different injections, in which the last four are used to ensure the standard deviation of the sample results remain within the instrument’s uncertainty ranges. The sample results from the LGR analyzer are calculated by comparing the sample set with standards using the standard notation of \[\delta\text{ sample} = \left(\frac{R_{\text{sample}}}{R_{\text{standard}}} - 1\right) \times 10^3\], where R is the \(^{18}\text{O}/^{16}\text{O}\) or \(^2\text{H}/^1\text{H}\) in the sample or standard. Additionally, deuterium excess (δ) was calculated using \[\delta\text{ (‰)} = \delta\text{D} - 8 \times \delta\text{¹⁸O}\] following Dansgaard (1964).

### 3.4.5 Terrain analysis

A 1 m DEM produced from 2009 stereo GEOEYE-2 satellite imagery (Collingwood, 2014) was used to calculate slope, aspect, flow points, flow accumulation and wetness index (WI) values for the surrounding landscape with Whitebox v3.2 (Lindsay, 2014). This WI raster was exported to ArcGIS and WI values adjacent to the river were extracted to generate separate longitudinal profiles on the east and west banks, respectively. Longitudinal WI data was then spatially binned to 10m intervals in order to better compare inferred inflows to the WI profiles and for regression tree analysis.

### 3.5 Results

#### 3.5.1 Meteorology, river discharge and soil temperature

Compared to previous seasons at the CBWAO, weather conditions during the 2014 melt season were relatively cool and dry (Figure 3.2). Mean air temperature for June and July were 0.0°C and 3.7°C, respectively, and total rainfall was 1 mm and 69.2 mm for the respective
months (Figure 3.2). Mean air temperatures for this season were colder than average and similar to 2003, while total precipitation for the 2014 season was similar to warmer years such as 2009 (Favaro and Lamoureux, 2014). Towards the latter part of the study period, air temperatures dropped below 0°C on July 19 and July 23 (Figure 3.2).

River discharge varied with solar radiation and air temperatures on a diurnal basis and overall discharge was dependent on the availability of snow for melt. Previous measurement has shown that measured discharge increases downstream due to contributions from incoming water sources into the West River channel throughout a summer season (Veillete, 2010). Peak flow for the 2014 melt season began shortly after temperatures increased above 0°C, and high flow lasted from June 17 to 22, with peak discharge reaching 4.2 m$^3$/s. Recession followed the nival melt and baseflow began after June 30. The lowest flow occurred on July 19 (0.07 m$^3$/s, depth 0.05 m, width 3.4 m). Discharge was lowest at approximately 1000h and higher in the late afternoon due to solar input that generated a regular pattern of daily meltwater runoff. Major rainfall events that rejuvenated discharge occurred on July 15, 24 and 27 coincided with colder air temperatures (Figure 3.2).

Soil temperatures for June averaged 1.2, -0.7, -3.5 and -5.7 °C at soil depths 5, 15, 40 and 65 cm, respectively. Mean July soil temperatures were 5.3, 4.1, 1.1 °C, and -0.9 for the same soil depths, respectively. Marked diurnal variations of soil temperature occurred at depths of 5 and 15 cm, while small variations in soil temperature at 40 and 65 cm were evident by early July (Figure 3.2).
Figure 3.2. West River discharge and precipitation during the 2014 summer season (A). A comparison of air temperature and West River temperature at different stations throughout the season (B). Soil temperatures at Lower Goose station (C). The shaded area represents the river temperature sampling interval. The grey interval indicates the period during which river temperature surveys and water sampling was carried out.
3.5.2 Seasonal river temperature

Prior to July 6, river temperatures were colder and less thermally variable than air temperatures. This pattern reversed after July 6, with air temperatures becoming cooler than river temperatures. Water temperature at station 11 (the most downstream station; Figure 3.1) and air temperature were significantly correlated after July 6 ($r = 0.71$, $n=528$, $p < 0.0001$). Both air and river temperatures were relatively synchronous, with river temperature showing a lag of approximately two hours compared to air temperature. Strong diurnal cycles were evident in both air and river temperatures throughout the season. The warmest temperatures were recorded in the upper reach of the West River, with stations 4 and 7 reaching a mean of 6.0°C for July, while stations 9 and 11 in the lower reach had colder temperatures during the same period with temperatures reaching 5.3°C and 5.4°C, respectively.

3.5.3 Longitudinal river temperature

A large zone of persistent channel snow separated the study portion of the river into an upper and lower reach during the 2014 summer season and notably impacted the downstream thermal conditions. Specifically, channel snow persisted at two locations at distances between ~1800-1700 m and ~1200-750 m upstream of the West Lake (Figures 3.1, 3.3, 3.4).
The morning longitudinal profiles (0800-1000 h) show an overall downstream warming trend in both reaches on July 6 and 10. On July 19, a persistent downstream cooling trend was measured in the upper reach and a warming trend occurred in the lower reach (Figure 3.3). Additionally, with the exception of July 19, the upper reach showed reduced downstream thermal gains (°C/km) as measured during the morning compared to the lower reach (Table 3.1).
Table 3.1. Summary of morning gauging station river discharge, net downstream temperature gradient, and net downstream water isotope change for the upper and lower reaches. Temperature changes reflect the difference between stations 6 and 7 in the upper reach and 9 and 11 in the lower reach. Isotopic values indicate the net downstream change in the respective reach for the same station intervals.

<table>
<thead>
<tr>
<th>Morning sampling</th>
<th>July 6</th>
<th>July 10</th>
<th>July 19</th>
<th>July 24</th>
</tr>
</thead>
<tbody>
<tr>
<td>Q (m³/s)</td>
<td>0.13</td>
<td>0.12</td>
<td>0.07</td>
<td>0.15</td>
</tr>
<tr>
<td>Upper reach (°C/1000m)</td>
<td>0.52</td>
<td>0.74</td>
<td>-0.16</td>
<td>-</td>
</tr>
<tr>
<td>Lower Reach (°C/1000m)</td>
<td>1.27</td>
<td>1.06</td>
<td>0.45</td>
<td>-</td>
</tr>
<tr>
<td>Upper reach (ΔδD ‰)</td>
<td>1.09</td>
<td>0.45</td>
<td>-3.46</td>
<td>4.77</td>
</tr>
<tr>
<td>Lower Reach (ΔδD ‰)</td>
<td>0.74</td>
<td>0.92</td>
<td>-0.38</td>
<td>-0.42</td>
</tr>
<tr>
<td>Upper reach (Δδ¹⁸O ‰)</td>
<td>0.09</td>
<td>0.16</td>
<td>-0.68</td>
<td>0.95</td>
</tr>
<tr>
<td>Lower Reach (Δδ¹⁸O ‰)</td>
<td>-0.01</td>
<td>0.30</td>
<td>-0.18</td>
<td>-0.29</td>
</tr>
</tbody>
</table>

Downstream thermal changes for both reaches were measured between stations 6 and 7 (upper reach) and 9 and 11 (lower reach) in order to reflect changes with no channel snow influence. The morning profiles representing no channel snow influence showed greater net thermal gains for the lower reach and the gradients were 1.27, 1.06, 0.45 °C/km on July 6, 10 and 19, respectively. By comparison, the upper reach had smaller thermal gains with gradients of 0.52 and 0.74 °C/km for July 6 and 10, and -0.16 °C/km on July 19 (Table 3.1). By including the thermal influence of residual channel snow, temperature fluctuations were larger compared to locations without snow influence. The measured 2.1 km reach (containing both the upper and lower reaches) with channel snow influence resulted in overall downstream temperature variance of the West River with gradients of 1.81°C, 1.57°C and 1.36°C for July 6, 10 and 19, respectively.

During the morning surveys, July 6 was characterized by numerous abrupt temperature decreases, while July 10 had a number of abrupt increases in localized areas along the river.
The morning of July 19, however, was characterized by more subtle variations of warming and cooling in both reaches.

![Figure 3.3](image)

**Figure 3.4** Longitudinal afternoon temperature profiles for July 6th, 10th, 16th and 19th and the DEM-derived wetness index (TWI) longitudinal profile for the east and west banks of the West River. Water sampling locations are also indicated (station 4-11). Longitudinal distance is measured upstream relative to the West Lake.

The afternoon (1600-1800 h) longitudinal temperature profiles reflected a similar pattern to the morning, with an overall downstream warming trend for all measured days in July across both reaches (Figure 3.4). The afternoon profiles for intervals without channel snow influence (station 6 to 7 and 9 to 11) showed comparable trends to the morning, but with greater overall thermal gains in both the upper reach (0.84, 0.72, 0.27, and 0.22 °C/km) and lower reaches (1.69, 0.73, 0.54, and 0.41 °C/km) for July 6, 10, 16 and 19, respectively (Table 3.2). The measured 2.1
km reach influenced by residual channel snow resulted in even greater thermal variance in the afternoon compared to the morning. The overall downstream temperature variance for the full 2.1 km reach containing residual channel snow influence for the West River resulted in gradients of 5.27°C, 2.56°C, 0.81°C, and 1.25°C for the afternoons of July 6, 10, 16 and 19, respectively.

The frequency and magnitude of abrupt variations in river temperatures was higher in the afternoon compared to the morning. Additionally, downstream temperature variations were substantially altered during the period of increased discharge that arose from rainfall on July 16, and were marked by reduced temperature variability and a dampened thermal influence of residual channel snow. As a result, the July 16 temperature profile exhibited the least amount of downstream thermal change in both reaches compared to the other dates measured.

<table>
<thead>
<tr>
<th>Afternoon</th>
<th>July 6</th>
<th>July 10</th>
<th>July 13</th>
<th>July 16</th>
<th>July 19</th>
<th>July 22</th>
<th>July 24</th>
</tr>
</thead>
<tbody>
<tr>
<td>Q (m³/s)</td>
<td>0.21</td>
<td>0.29</td>
<td>0.26</td>
<td>0.2</td>
<td>0.11</td>
<td>0.19</td>
<td>0.22</td>
</tr>
<tr>
<td>Upper reach (°C/1000m)</td>
<td>0.84</td>
<td>0.72</td>
<td>-</td>
<td>0.27</td>
<td>0.22</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Lower Reach (°C/1000m)</td>
<td>1.69</td>
<td>0.73</td>
<td>-</td>
<td>0.54</td>
<td>0.41</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Upper reach (ΔδD ‰)</td>
<td>3.79</td>
<td>-2.25</td>
<td>0.76</td>
<td>1.19</td>
<td>-1.44</td>
<td>0.58</td>
<td>-1.92</td>
</tr>
<tr>
<td>Lower Reach (ΔδD ‰)</td>
<td>1.05</td>
<td>0.62</td>
<td>1.29</td>
<td>3.56</td>
<td>1.51</td>
<td>-0.57</td>
<td>0.54</td>
</tr>
<tr>
<td>Upper reach (Δδ¹⁸O ‰)</td>
<td>0.48</td>
<td>-0.03</td>
<td>0.27</td>
<td>0.28</td>
<td>0.17</td>
<td>-0.08</td>
<td>-0.43</td>
</tr>
<tr>
<td>Lower Reach (Δδ¹⁸O ‰)</td>
<td>-0.13</td>
<td>0.15</td>
<td>0.25</td>
<td>0.20</td>
<td>0.33</td>
<td>0.12</td>
<td>0.06</td>
</tr>
</tbody>
</table>

Table 3.2. Summary of afternoon gauging station river discharge, net downstream temperature gradient, and net downstream water isotope change for the upper and lower study reaches. Temperature changes are indicated in snow-free channel intervals, and reflect difference between stations 6 and 7 in the upper reach and 9 and 11 in the lower reach. Isotopic values indicate the net downstream change in the respective reach for the same station intervals.
3.5.4 Inferred channel inflows

Early during the 2014 season, analysis of cumulative departure ($\Delta T_i$) revealed that known surface inflows from Ptarmigan (PT), ALD05 (AL), Goose (GS) and Muskox (MX) creeks were evident as localized (and longitudinally abrupt) inputs of warmer water into the West River at their inflow points. Longitudinal measurements of temperature revealed other localized fluxes of warm water in addition to these known point sources. Based on temperature change thresholds $\Delta T$ exceeding $+0.01^\circ$C, inferred surface inflows which exhibited warm fluxes were delineated for July 10, 16 (Figure 3.5), and 19 (Figures 3.6, 3.7). As the season progressed, observed surface inflows either ceased or contributed fewer instances of measurable thermal changes in the river. Results below emphasize a comparison between July 10 and 19, representing the widest range of conditions in this study.

Temperature survey ($\Delta T$) measurements also demonstrate two types of cold water input responses along the river. Areas with residual channel snow generated inputs of cold water followed by a downstream interval of increased temperature variability. By contrast, locations without channel snow or snow in proximity (upslope) of the river exhibited initial source points of cold water input, with a more localized and subtle downstream thermal mixing zone. These cold water inflows were used to delineate inferred subsurface inflows for July 10 (Appendix A), 16 (Figure 3.5), and 19 (Figures 3.6, 3.7) based on measurements of $\Delta T$ exceeding a -$0.01^\circ$C threshold. The most notable area in which cold subsurface inflows were observed consistently throughout the season was located between 300-375 m at station Beta and also between 1240-1260 m upstream of the Muskox Creek inflow (Figure 3.1). Compared to other point source inputs of cold water, those associated with channel snow were characterized by a greater initial temperature decrease and a longer interval of thermal mixing along the river channel earlier
during the season such as July 10 (Appendix A), and lessened in intensity and range along the river channel as the season progressed as shown on July 19 (Figures 3.6, 3.7).

Source point inputs of inferred cold subsurface inflows were evident as much more subtle fluctuations in river ΔT compared to source point inputs of channel snow. Precise location of these point sources was also confounded by the presence of known and inferred surface inflows, particularly during the morning and afternoon of July 10 (Appendix A). A total of 41 surface and 28 subsurface inflows in the morning, and 59 surface and 44 subsurface inflows in the afternoon were delineated for July 10. By comparison, the river survey on July 19 indicated 29 surface and 21 subsurface inflows in the morning, and 26 surface and 24 subsurface inflows in the afternoon (Figures 3.6, 3.7).

Intraday results on July 10 indicate that a larger number of ΔT exceeded ±0.01°C thresholds in the afternoon compared to the morning, which suggests that inferred surface and subsurface inflows are more active during the afternoon. However, the magnitude of measured ΔT for warm and cold water inputs for July 10 (Appendix A) is greater in the morning compared to the afternoon, and a similar pattern was observed during the morning (Figure 3.6) and the afternoon (Figure 3.7) of July 19. By contrast, temperature departures on the morning of July 19 indicate that subsurface inflows had a greater downstream influence on the river than surface inflows, which contributed to overall downstream cooling of the upper reach. Downstream warming of the upper reach in the afternoon of July 19 suggest that inferred surface inflows became more active and influential compared to subsurface inflow as the peak hydrological activity was reached during the diurnal cycle for the day.

Cumulative downstream ΔT departures and temperature threshold exceedances measured during runoff associated with July 16 rainfall (Figure 3.5) suggests known and inferred surface
(warmer) inflows were widespread and had greater thermal contributions to the West River compared to inferred subsurface (colder) inflows. Additionally, known surface inflows which had become less active during the season were rejuvenated by the rainfall, as evidenced by well-defined positive temperature offsets at catchment inflow points PT, AL, GS and MX (Figure 3.5).

Figure 3.5 Comparison of ΔT and the cumulative departure of ΔT on the afternoon of July 16 with known tributary inflows of PT, AL, GS and MX represented by arrows. Q represents the mean discharge during the temperature survey at the West River station (Figure 3.1). Surface, and subsurface inputs relative to the river are shown based on ±0.01°C ΔT threshold exceedances. Distance is measured upstream relative to West Lake. Insets show detailed longitudinal temperature changes for two intervals of interest. Interval AB indicates the impact of residual channel snowpack, and interval CD indicates the impact of a surface water inflow.
Figure 3.6 Comparison of ΔT and the cumulative departure of ΔT on the morning of July 19 with known tributary inflows of PT, AL, GS and MX represented by arrows. Q represents the mean discharge during the temperature survey at the West River station (Figure 3.1). Surface, and subsurface inputs relative to the river are shown based on ±0.01°C ΔT threshold exceedances. Distance is measured upstream relative to West Lake. Insets show detailed longitudinal temperature changes for two intervals of interest. Interval AB indicates the impact of residual channel snowpack, and interval CD indicates the impact of a surface water inflow. Note the diminished effect of snowmelt influence in interval AB.
Figure 3.7 Comparison of ΔT and the cumulative departure of ΔT on the afternoon of July 19 with known tributary inflows of PT, AL, GS and MX represented by arrows. Q represents the mean discharge during the temperature survey at the West River station (Figure 3.1). Surface, and subsurface inputs relative to the river are shown based on ±0.01°C ΔT threshold exceedances. Distance is measured upstream relative to West Lake. Insets show detailed longitudinal temperature changes for two intervals of interest. Interval AB indicates the impact of residual channel snowpack, and interval CD indicates the impact of a surface water inflow. Note the diminished effect of snowmelt inflow in interval AB.

3.5.5 Seasonal downstream water stable isotopic composition

Isotopic composition of all samples in the river fall to the right of the Global Meteoric Water Line (GMWL) and suggests that water from the 2014 summer season was subject to varying degrees of evaporative enrichment (Figure 3.8). Isotopic composition indicates that the West River is initially more depleted at the beginning of the season and becomes progressively enriched for both the upper and lower reaches as the season continues (Figure 3.8). Sample δD
ranged from -184.7 to -168.0‰ in the morning and from -186.0 to -171.5‰ in the afternoon throughout the season. Corresponding δ¹⁸O from the same sample points ranged -24.3 to -21.6‰ in the morning and -24.3 to -22.3‰ in the afternoon. Enrichment of δD relative to δ¹⁸O on July 16th followed by a subsequent depletion on July 19 and 22 suggest that precipitation between July 13-16 delivered relatively enriched water to the river. While no rain or permafrost data was available for 2014, 2012 rainfall at CBAWO ranged from -151.1 to -121.8‰ for δD and -19.6 to -15.5‰ for δ¹⁸O. Similarly, ice from a 2012 permafrost core ranged from -139.8 to -125.0‰ for δD and -17.9 to -16.2‰ for δ¹⁸O. Hence, both rainfall and permafrost data from 2012 were isotopically enriched relative to the West River in 2014. Overall isotopic variation of the river between all sample stations for δD was lowest on July 6 and 10, and showed increased variability between sample stations in the morning of July 19 and July 24 compared to the afternoon of the same days (Figure 3.9).
Figure 3.8: Isotopic composition of water sampled from the West River and catchment tributary samples in 2014. West River samples shown are from stations 4 to 11 during the afternoon. Catchment samples from tributary creeks PT, AL, GS and MX from the same sample dates are also shown. The detailed panel provides additional detail for clustered samples. Permafrost from one core and rain data from 2012 is indicated to illustrate the isotopic composition of water sources. The Global Meteoric Water Line (Gibson et al., 1993) is also shown for reference. All samples indicate analytical uncertainties, although these are often less than the symbol size and not apparent.

Intraday δD values revealed there was overall downstream enrichment on each morning sampled (Figure 3.10) and in the afternoon as well (Appendix C) throughout the season. While an overall enrichment trend for the river is evident, there is considerable isotopic variability within the upper and lower reaches. Isotopic and channel inflow analysis (previous section)
suggest that known and inferred surface inflows contribute to the observed pattern of downstream isotopic enrichment (Figures 3.8, 3.9), although the coarse spatial resolution of isotopic sampling limits interpretations. Tributary inflows from catchments PT, AL, AD and MX had strongly enriched isotopic composition relative to the river, with δD ranging between -172.2 to -139.8‰ and δ18O between -22.1 to -17.5‰ (Figure 3.8). Additionally, intervals which have evidence for inferred subsurface inflows based on temperature analysis also correspond to downstream locations that show isotopic depletion on July 19 in the morning and afternoon.

Therefore, δD and δ18O broadly corroborate the temperature data that indicated a net downstream increase in temperature due to surface inflows also show more increased isotopic enrichment in the corresponding reach (Table 3.2), while instances where a net loss in downstream temperature occurred due to inferred subsurface inflows (July 19, morning), particularly during the late season, showed isotopic depletion (or reduced downstream enrichment) over a specific reach (Table 3.1). However, a finer isotopic sampling resolution is required for more conclusive results.
Figure 3.9: Seasonal progression of isotope composition (δD) for West River, ranging from stations 4 to 11, during the 2014 season between A) morning and B) afternoon.

Locations that contained residual channel snow also similarly corresponded to observed downstream δD depletion, but were only evident earlier during the season (July 6 and 10) at the beginning of the lower reach (Figure 3.8). Samples collected on July 16 (Figures 3.8, 3.9) had enriched δD and δ¹⁸O across both reaches, indicating a pattern of enrichment in response to rainfall runoff events. Another rainfall event on July 24 also generated isotopic enrichment and increased intra-station variability (Figure 3.9). Overall, isotopic composition appeared to be
depleted at the occurrence of clusters of inferred subsurface inflows, while channel snow did not appear to result in a clear downstream depletion signature based on this coarse sampling resolution. During the morning, the net downstream isotopic change in the upper reach was 1.09, 0.45, -3.46, and 4.77 ‰ δD/km on July 6, 10, 19, and 24, respectively. In comparison, the upper reach had respective temperature changes of 0.52 and 0.74 °C/km for July 6 and 10, while a net thermal loss of -0.16 °C/km on July 19 (Table 3.1).

Figure 3.10: Longitudinal variations in δD sampled at stations 4 to 11 (Figure 3.1) during the morning. July 19 morning inferred inflows and channel snow are shown for reference.
3.5.6 Terrain wetness index and inferred inflows

Topographic wetness index (TWI) longitudinal profiles for both the east and west banks of the West River were generated by using a digital elevation model to further investigate source point locations of inferred surface and subsurface flow based on several factors such as elevation, aspect and slope. Index values from each bank were generated separately to reflect differences in the respective slope and drainage patterns.

Analysis of wetness profiles in conjunction with thermal evidence for surface and subsurface inflows suggest that elevation, slope and aspect can be used to localize hydrological flow paths that explain the observed thermal variability in the West River. Results indicate that topographic controls of both the east and west banks in proximity to the West River generate highly variable surface and subsurface contributing areas and potential inflow pathways. Tributaries located in the east bank of the upper reach have elongated flow pathways that reach the catchment interfluve at ~80-90 m asl, while the pathways on the west bank of the upper reach have shorter and more limited contributing areas due to a lower elevation intermediate upslope interfluve (Figure 3.11). Broadly, this pattern is reversed for the lower reach, where the east bank has more limited flow pathways due to limited slope length, and most of the west bank also exhibits wetness index characteristics conducive to surface and subsurface flow pathways (Figure 3.11).
Figure 3.11: TWI for the study reaches of the West River for the east and west banks. Also shown is ΔT and inferred source inflows on the morning of July 19. The contributing bank of surface and subsurface inflows is inferred on TWI levels. Note the inverted scale for the east bank TWI data.

Comparison of inferred inflows and the TWI records frequently shows a close association between a TWI maximum value on one bank and an inflow. For example, on the morning of July 19, results indicate that the TWI maxima-inflow associations are more frequent on the west bank in the upper reach of the river for subsurface flow, while the TWI of the east bank is more conducive to specific surface inflows for the same reach, represented by greater proportions of known and inferred surface inflows for the East Bank (Figure 3.11). For the lower reach on the same survey date, the majority of inferred surface and subsurface inflows are associated with TWI maxima on the east bank, while the west bank has surface inflows limited to the channel 125-500 m from West Lake (Figure 3.11). Regression tree analysis indicates that the distribution
of channel inflows compared to maximum wetness index values in 10 m intervals and ΔT can be attributed to three major categories (Figure 3.12): 1) low TWI and cold temperature changes; 2) high TWI and cold temperature changes; and 3) high TWI, warm temperature changes. For the morning of July 19, residual channel snow is associated with category 1 and surface inflows are associated with category 3. Inferred subsurface inflows are more variable and are mainly associated with mid- to high TWI values (7 to 9.5). Additionally, several of the subsurface inflows exhibiting the largest temperature departures were associated with higher TWI values in comparison to other inflows of the same type (Figure 3.12). Similar patterns were discerned with respect to surface inflow pathways on the afternoons of July 16 and 19 (Appendix E).
Figure 3.12: Temperature change (ΔT) and maximum wetness index compared in a regression tree analysis for the morning of July 19 inferred inflows in the study reaches. The range of respective surface, subsurface inflows and channel snow channel inflows are delineated by boxes, forming three main response zones.
3.6 Discussion

3.6.1 Thermal regime for a river in the Canadian High Arctic

Water exchanges and the thermal regime of rivers have been investigated in a limited way in arctic settings, particularly in light of changing permafrost conditions. Similar studies have shown that there are multiple controls to the thermal dynamics of rivers and streams in Arctic (Blaen et al. 2012), Alpine (Uehlinger et al. 2003; Brown et al. 2006, 2008; Dickson et al. 2012) and temperate regions (Webb and Zhang, 1999; Westhoff et al., 2011), which affect river temperature spatially and temporally that Cassie (2006) summarized as: 1) short term diel and seasonal atmospheric conditions; 2) stream discharge; 3) topography; and 4) streambed interactions such as hyporheic exchange and groundwater inputs (Figure 3.13).

![Figure 3.13: Categories which influence the thermal regime of rivers in Arctic systems (modified for Arctic settings, from Cassie, 2006).]
Incoming solar radiation is lower at high latitudes and causes less thermal variability on a
diurnal scale in comparison to alpine systems (Blaen et al. 2012). Clear diel differences in
temperature are evident in the West River which are controlled primarily by solar radiative heat
gains (Cassie, 2006; Webb and Zhang, 1999; Webb et al., 2008), and are marked by relatively
synchronous intraday variations of air and water temperatures. As such, variability in
downstream warming is heavily dependent on solar irradiance and meteorological controls such
as surface albedo and cloud cover (Uehlinger et al. 2003; Cassie 2006; Webb et al., 2008; Blaen
et al. 2012). In temperate and some alpine regions, the vegetation canopy often strongly affects
river temperatures through shading effects (Johnson and Jones, 2000; Mellina et al., 2002;
Moore, 2006; Webb et al., 2008). However, no significant vegetation shading occurs in the high
arctic due to the low tundra vegetation (Figure 3.14).

Areas of consistent thermal variability and strong thermal mixing between 1800-1700 m
and 1200-750 m upstream of West Lake, particularly on July 6 (Figure 3.3), were likely related
to the input of cold meltwater from residual channel snow (Webb and Nobilis, 1994; Brown et
al., 2006; Blaen et al., 2012), which persisted in these areas throughout the season (Figure 3.14).
In this case, channel snow in close proximity to the river channel was associated with a
prominent downstream cooling effect compared to an overall pattern of downstream temperature
gain. The correlation between air and water temperatures was weaker when the river was
dominated by channel snow prior to July 6, and indicates that the presence and proximity of
channel snow to river channels is an important component for the river heat budget in this arctic
setting, particularly in the early runoff season. Although channel snow contributions are
significant, the magnitude of these contributions is highly dependent on atmospheric conditions
on a daily scale, where increases in solar radiation and sensible energy gains from warming air temperatures provide greater localized inputs of cold meltwater (Woo, 2012).

Small rivers and streams are strongly affected by atmospheric forcing due to relatively small discharge (Q) quantities which require less thermal energy to impart warming (Gu and Li, 2002). Downstream warming of the West River in this arctic system is consistent with river systems of similar sizes elsewhere, which have a downstream warming rate of $\sim 0.6^\circ C \text{ km}^{-1}$, compared to larger rivers that exhibit typically lower rates ($\sim 0.09^\circ C \text{ km}^{-1}$) (Zwieniecki and Newton, 1999; Torgerson et al., 2001; Caissie, 2006).

While rainfall was limited and episodic throughout the season, these events coincided with dampened air and water temperatures across both reaches. However, such small precipitation inputs are not likely to have provided strong temperature fluctuations directly from precipitation (Webb and Zhang, 1999; Blaen et al. 2012). The findings from this study suggest that precipitation provided advective energy inputs sourced from surface and near surface hillslope pathways (Brown and Hannah, 2007), which provided the observed thermal changes to the river. Additional variables can be considered when looking at thermal dynamics and atmospheric forcing in Arctic rivers, such as wind speed, humidity, and evaporation, although these exchanges appear to be small compared to the dominant influence of solar heating and snow melt contributions.
These external energy components are important for characterizing the seasonal and broad patterns of river temperature, but temperature differences measured over short distances and time intervals reflect different controls. While atmospheric conditions strongly impact stream discharge variables, topography acts as a routing factor that influences surface and subsurface inflows, which considerably affect river temperatures (Cassie, 2006; Blaen 2012; Dingman 2012; Woo 2012). Hence, the heat budget at a given location or reach of an arctic river can be considered to be the discharge-weighted sum of thermal exchanges as follows:

\[ T_n = \uparrow T_r \pm \downarrow T_e \pm \uparrow T_a \pm \uparrow T_p \pm \uparrow T_s(Q_s) \pm \downarrow T_{ss}(Q_{ss}) \pm \uparrow T_{ch}(Q_{ch}) \pm \downarrow T_{hyp}(Q_{hyp}) \]  

(3.4)

where \( T \) = temperature effect of the various components: \( Q \) = total discharge, \( r \) = net radiation, \( e \) = evaporation and condensation, \( a \) = air temperature, \( p \) = precipitation, \( s \) = surface inflows, \( ss \) = subsurface inflows, \( ch \) = channel snow, and \( hyp \) = hyporheic exchange.

In the above formulation, groundwater is not taken into account as continuous permafrost likely prevents deeper groundwater from interacting with surface water which would otherwise...
limit or buffer river temperature variability (Brown et al., 2005; Zarnetske et al., 2007; Blaen et al., 2012). Notably, climate change, particularly in areas of thin or discontinuous permafrost may result in groundwater interactions with the surface (AMAP, 2012).

The temperature of some of these components can cause the net temperature of a river to undergo cyclical thermal stages on a seasonal and daily timescale. For example, increases of incoming solar radiation (r) and air temperature (a) cause overall river temperature to increase during the early season, stabilize during the mid-season, and decrease in the late season as the seasonal influence of solar radiation and air temperature decreases (Liu et al., 2005, Caissie, 2006; Blaen et al., 2012). Over a 24-hour period, river temperatures also exhibit similar fluctuations in thermal conditions where temperatures increase during the late morning, peak during afternoon, and decrease during the late evening (Webb and Zhang, 1999; Brown et al., 2005; Caissie 2006).

Over short distances, stream discharge and streambed interactions such as surface inflows (s), subsurface inflows (ss), hyporheic exchange (hyp), and channel snow (ch) are potentially more thermally influential compared to evaporation and transpiration or precipitation processes (Webb and Zhang, 1999; Gu and Li, 2002; Caissie; 2006; Brown and Hannah, 2007). Results from longitudinal temperature profiles (Figures 3.5, 3.6, and 3.7) demonstrate these localized thermal changes while atmospheric conditions are responsible for broader thermal changes across longer distances in the reach of a river. Hence, this research demonstrates the important role individual hydrological sources and pathways contribute to fine-scale temperature variations in and along an arctic river.
3.6.2 Channel inflows and stream discharge conditions

At a finer spatial scale, mixing of surface inflows and subsurface inflows in river systems causes localized changes in water temperature, as evidenced from West River temperature variations in this study (Figures 3.5, 3.6, 3.7). Specifically, surface inflows are indicative of warmer localized mixing relative to river temperature and contribute to net downstream increases of river temperature (Uehlinger et al., 2003; Caissie, 2006; Webb et al., 2008). The magnitude of localized warming provided by these surface inflows are highly dependent on inflow volume and atmospheric conditions. As is the case with rivers, surface inflow temperatures can vary widely based on diel and seasonal fluctuations in atmospheric conditions. Streams which have smaller discharge volumes are more sensitive to these atmospheric fluctuations, as small and shallow channels tend to be highly exposed to meteorological conditions and less overall energy is required to warm smaller quantities of water (Mosley, 1983; Caissie, 2006).

While changes in surface water temperatures are mainly caused by thermal energy exchanges between water and the atmosphere (Webb and Zhang, 1999; Johnson, 2004; Caissie, 2006; Carrivick et al., 2012), subsurface flows in permafrost settings are impacted by the presence of a frost table throughout the runoff season. Soil temperature and pore water become cooler with depth due to slow conductive heat transfer from the surface (Woo, 2012). This soil thermal profile results in subsurface inflow temperatures being cooler relative to surface inflows as well as the West River. Hence, while subsurface inflows provide localized input of cooler water, the observed thermal effects of these inflows are more subtle, which indicates that subsurface discharge ($Q_{ss}$) to the river is correspondingly smaller than comparable surface inflows (Figures 3.6, 3.7, 3.8). Moreover, the soil temperature profile will sustain relatively cold subsurface inflows regardless of how deep the active layer is in a given year. Hence, the cold
thermal influence of subsurface inflows can be expected to occur throughout the season and in a wide range of interannual climatic conditions.

While subsurface discharge inflows appear to be volumetrically limited, periods of lower river discharge cause subsurface contributions to become relatively more significant, so much so that the normal downstream warming evident in a wide range of river settings changes to downstream cooling. This effect was most evident on the morning of July 19 (Figure 3.7) in the upper reach, and indicates that atmospheric and surface thermal exchanges were less than subsurface contributions. Low river discharge that morning allowed for subsurface inflows to be proportionately more influential in terms of relative water contributions, with a measurable impact on the thermal regime of the West River. Heat loss to the atmosphere cannot explain this pattern, as the lower reach was characterized by downstream warming during the same sampling interval (Figure 3.7). Hence, this differential thermal response between the reaches suggests that the upper reach has a greater relative contribution from subsurface water. While the impact of subsurface inflows is most apparent in the upper reach during the low flow conditions on the morning of July 19, the net downstream temperature gradients for earlier dates with higher discharge also show reduced thermal gains compared to the lower reach (Tables 3.1, 3.2) and suggests that while the effect is most evident at lowest flow, cool subsurface flows were likely making a measurable impact on river water temperature throughout the July study period.

The magnitude of downstream cooling of the West River suggests that surface inflow contributions are more prominent early during the season, and tend to have greater thermal influence compared to subsurface inflows due to atmospheric conditions and discharge volume. This dynamic is further evidenced through greater temperature increases at locations with surface
inflows early during the season, while subsurface inflow contributions show smaller negative
temperature fluctuations later in the season when river discharge is lower.

While these results have been considered in terms of surface and subsurface inflows to
the river, the role played by hyporheic exchanges should also be considered (Williams and
Smith, 1989; Chapin et al., 1992; Zarnetske et al., 2007; Westhoff et al., 2011). It is important to
note that it is often difficult to determine the flux, volume and location of the hyporheic zones in
a channel system (Westhoff et al., 2011). Due to the sampling limitations of this study, instances
of hyporheic exchange were not identified or quantified, although its potential role in the
observed thermal dynamics is acknowledged. Pools and riffle sequences are present in both the
upper and lower reaches of the West River where hyporheic exchanges are broadly known to
occur (Boulton et al., 1998; Zarnetske et al., 2007; Westhoff et al., 2011). Indeed, what is
inferred to be thermal influences related to lateral subsurface inflows could be related to in part
to hyporheic exchange. Further investigation to distinguish these two pathways is required.

However, limited available stable isotope data suggests some downstream changes in
isotopic composition consistent with inflows rather than thermal alteration of channel water
through hyporheic routing. The clear isotopic enrichment of tributary catchments in comparison
to the isotopic composition of the West River (Figure 3.8) is likely due to the increased
importance of evaporation due to lower discharge, surface ponding and soil water evaporation.
Precipitation events rejuvenated various surface inflows to the West River, which provided water
that was warmer than the West River (Figure 3.5) and contributed towards downstream isotopic
enrichment on July 16 (Figure 3.9, Table 3.2). Isotopic data from 2012 also corroborates this
assessment, with precipitation measured to be more enriched than the West River and its
tributaries (Figure 3.8). Instances of downstream isotopic depletion are more apparent later in the
season, when discharge is lowest and presumably less significant volumetrically compared to subsurface inflows, due to older soil water being released by permafrost thaw and directed by subsurface pathways into the river (Gazis and Feng, 2004; Darling et al., 2006). For example, the upper reach on the morning of July 19 contains numerous inferred subsurface inflows and was characterized by both downstream cooling and isotopic depletion (Figure 3.10). Reduced downstream isotopic depletion during in afternoon (Table 3.2) during slightly higher discharge further demonstrates that the relative influence of surface and subsurface inflows over river isotopic composition are dependent on relative discharge water contributions, and argue that the thermal observations are consistent with water inflows to the channel rather than hyporheic exchanges. Limited ground ice data, collected at Cape Bounty in 2012 (Lamhonwah, unpublished data) suggests that the isotopic composition of ground ice is substantially enriched relative to the West River and cannot explain the observed downstream depletion. However, the isotopic composition of various types of ground ice across the Canadian Arctic generally have δ¹⁸O signatures of -26 to -36‰ and δD ranging between -200 to -255‰ (Michel, 2011), which could account for downstream depletion in the West River although further ground ice data would be necessary to explain the observed isotopic pattern.

3.6.3 Topographic influence and watershed structure on flow pathways

Topographic factors such as watershed morphology and hillslope drainage patterns influence the routing of inflows along a river and it has been found that during periods of high wetness, hillslopes are highly transmissive and help to direct significant quantities of water to streams, rivers and areas adjacent to hillslopes (Jencso et al., 2009). Broadly, the West River TWI (Figure 3.15) suggests this type of slope-based transmissivity, where the likelihood of
saturated soils increases at the base of hillslopes and in depressions on the landscape where there is convergence of both surface and subsurface flow (Barling et al., 1994). Specifically, the west bank of the upper reach (Figure 3.15) is an example of two processes regarding hillslope based hydrological transmissivity, where the west hillslope separating both reaches causes the convergence of channelized surface inflows north of the hillslope (location A) in contrast to the shorter slope along the west bank that limits convergent flow and channelization, and likely causes subsurface water inflows (location B).
Figure 3.15: A portion of the Cape Bounty Arctic Watershed Observatory (CBAWO) study site showing a DEM-derived topographic wetness index (TWI) map for the study reach of the West River. Inferred surface inflows, subsurface inflows and channel snow on the morning of July 19 are indicated. Enlarged panels (right) indicate the distribution of inflows between the upper and lower reaches in more detail.
Similarly in the lower reach, the longer hillslope drainage to the northeast of location C (Figure 3.15) causes convergence of channelized surface inflow and is similar to location A. While flow and slope length is limited, there is still some localized hydrological connectivity to the West River (Figure 3.16) through convergent slope soilwater flow that results in localized subsurface flow inputs (Anderson et al., 2009).

![Figure 3.16: Convergence of channelized surface inflow at hillslope location C (photo E) (July 13, 2014) and (photo F) (July 19, 2014) inflow point to the left of the West River of a limited incised stream network sourced from hillslope location C, in the lower reach.](image)

In the upper reach of the West River, major tributaries (PT, AL, GS and MX) have greater lengths of surface flow on the east bank, which also help to contribute to downstream increased stream temperature. In contrast, some locations with known large tributaries with incised channels do not lead to a marked thermal impact in the West River due to cessation or limited relative flow into the West River (Figure 3.15). As previously mentioned, solar radiation provides downstream warming to streams and rivers, which increases with distance (Cassie, 2006). The lower reach exhibits similar channelization characteristics on the west bank (location D) (Figure 3.15). Hence, while these conditions are highly specific to this location, they illustrate the role played by topography in contributing to inflow characteristics for both surface and subsurface waters, and provide a means to differentiate these contributions by reach.
3.7 Conclusion

This research provides insights into the processes that explain longitudinal temperature variations caused by surface and subsurface inflows to a small arctic river, as well as the landscape controls that affect these inflows. Projected changes of temperature and precipitation in high latitude regions will have an impact on these longitudinal temperature variations through a deepening active layer. Surface and (subsurface) inflows are identifiable by abrupt downstream warming and (cooling) at specific locations, and the additional effect of residual channel snowpack is also evident by zones of river water cooling. These thermal influences were evident throughout the baseflow period and muted overall downstream warming throughout the measurement period within one of the study reaches, which suggests a greater presence of subsurface inflow compared to another reach further downstream. Additionally, the effects of climate change will inevitably change the timing and influence of channel snowpack cooling seen in this study, which acts as a thermal refuge for aquatic organisms. Based on these thermal profiles, topography and a deepening active layer has a large influence on the channelization and localization of surface and subsurface flow to the West River.

These results demonstrate that thermal dynamics, in conjunction with isotopic tracers, can be used to track channel-slope water interactions in arctic hydrological systems, work previously limited to alpine and temperate settings. It advances our understanding of the impact of changing surface and subsurface inflow patterns within a season in a High Arctic setting and contributes to related research focused on hydrological heat budgets, ecosystem health and to help understand how hydrological change, thermal variations and changing water sources will affect aquatic ecosystems, particularly aquatic organisms sensitive to water temperature change.
Chapter 4

Conclusions and Future Work

River systems all around the world are influenced thermally; however changing climate causes new implications for aquatic ecosystems and overall thermal and isotopic dynamics of hydrological systems at high latitudes. This research, conducted at the Cape Bounty Arctic Watershed Observatory, related to hydrological processes and thermal dynamics can be summarized as follows:

1) Channel snow and solar radiation are the largest contributing factors to thermal patterns in river systems with a similar landscape to the CBAWO. Channel snow contributions diminish during the season and are less evident during low morning flow, while atmospheric conditions provide greater contributions in the afternoon and later in the season.

2) Channel snow and subsurface inflows provide net thermal cooling and depletion of δD and δ^{18}O to the river, while surface inflows provide net downstream temperature gains and enrichment of δD and δ^{18}O. Precipitation events provide varied results relating to thermal gains/losses, enrichment/depletion depending on active layer depth and the proportion of older subsurface water relating to the amount of precipitation. Throughout the season, subsurface inflows provide subtle contributions to the West River, while surface inflows show more pronounced thermal changes early in the season and diminish as meltwater availability is reduced. Discharge volumes between surface inflows, subsurface inflows and the river have a significant impact on relative thermal and isotopic variations occurring within a river system.
3) Topography has an affect towards the channelization of surface and subsurface inflows into the West River, and terrain analysis can be used to localize potential surface and subsurface inflow types and patterns. Specifically, topographical factors such as watershed structure and hillslopes influence the routing of channel inflows causing localized thermal changes and variations within a river system.

Future work following this type of project should have measurements on a more frequent intraday and seasonal timescale. Additional measurements such as electrical conductivity should also be considered for use, ideally with the same precision as the temperature sensor used in this study. Similarly, a more accurate GPS survey to better synchronize measurements with high quality DEMs would reduce spatial uncertainty of mapped inflows. Similarly, isotopic and chemical sampling should be performed at a finer spatial scale along the river channel and at known catchment inflows in order to identify subtle fluctuations in the river. Areas of interest should be studied at a finer scale, with thermal profiles being recorded in a grid pattern above and below the riverbed. Discharge values for known tributary inflows could provide more precise information on how various volumes of surface inflows influence river temperatures throughout an Arctic summer season.
References


Appendix A
Longitudinal Temperature Profiles

This appendix includes the remaining longitudinal temperature profiles for July 6 and July 10. Delta temperature (ΔT) thresholds for July 6 were not delineated due to heavy influence by channel snow.

Figure A-1: Comparison of ΔT and the cumulative departure of ΔT on the morning of July 10 with known tributary inflows of PT, AL, GS and MX represented by arrows. Q represents the mean discharge during the temperature survey at the West River station (Figure 3.1). Surface, and subsurface inputs relative to the river are shown based on ±0.01°C ΔT threshold exceedances. Distance is measured upstream relative to West Lake. Insets show detailed longitudinal temperature changes for two intervals of interest. Interval AB indicates the impact of residual channel snowpack, and interval CD indicates the impact of a surface water inflow.
Figure A-2: Comparison of $\Delta T$ and the cumulative departure of $\Delta T$ on the afternoon of July 10 with known tributary inflows of PT, AL, GS and MX represented by arrows. Q represents the mean discharge during the temperature survey at the West River station (Figure 3.1). Surface, and subsurface inputs relative to the river are shown based on $\pm 0.01^\circ C$ $\Delta T$ threshold exceedances. Distance is measured upstream relative to West Lake. Insets show detailed longitudinal temperature changes for two intervals of interest. Interval AB indicates the impact of residual channel snowpack, and interval CD indicates the impact of a surface water inflow.
**Figure A-3:** Comparison of $\Delta T$ and the cumulative departure of $\Delta T$ on the morning of July 6 with known tributary inflows of PT, AL, GS and MX represented by arrows. $Q$ represents the mean discharge during the temperature survey at the West River station (Figure 3.1). Surface, and subsurface inputs relative to the river are shown based on $\pm 0.01^\circ C$ $\Delta T$ threshold exceedances. Distance is measured upstream relative to West Lake. Insets show detailed longitudinal temperature changes for two intervals of interest. Interval AB indicates the impact of residual channel snowpack, and interval CD indicates the impact of a surface water inflow.
Figure A-4: Comparison of $\Delta T$ and the cumulative departure of $\Delta T$ on the afternoon of July 6 with known tributary inflows of PT, AL, GS and MX represented by arrows. $Q$ represents the mean discharge during the temperature survey at the West River station (Figure 3.1). Surface, and subsurface inputs relative to the river are shown based on $\pm 0.01^\circ C$ $\Delta T$ threshold exceedances. Distance is measured upstream relative to West Lake. Insets show detailed longitudinal temperature changes for two intervals of interest. Interval AB indicates the impact of residual channel snowpack, and interval CD indicates the impact of a surface water inflow.
Appendix B
River Temperature Distributions

This appendix section includes histograms for temperature distributions measured in the West River on the mornings and afternoons of July 6, 10, 16 and 19.

Figure B-8: Compilation of July 6th, 10th and 19th histogram of West River temperatures in the morning, coupled with normal distributions.
Figure B-9: Compilation of July 6\textsuperscript{th}, 10\textsuperscript{th}, 16\textsuperscript{th}, and 19\textsuperscript{th} histogram of West River temperatures in the afternoon, coupled with normal distributions.
Appendix C
Isotope Data

This appendix section includes the stable water isotopic profiles not included in chapter 3.

Figure C-1: Longitudinal $\delta^{18}O$ for the West River sampled at stations 4 to 11 (Figure 3.1) during the morning on July 6, 10, 19 and 24. Inferred inflows and channel snow indicated for July 19 morning for reference.
Figure C-2: Longitudinal δD for the West River sampled at stations 4 to 11 (Figure 3.1) during the afternoon on July 6, 10, 13, 16, 19, 22 and 24. Delineations for July 19th's afternoon inferred inflows and channel snow is also shown for reference.
Figure C-3: Longitudinal δ¹⁸O for the West River sampled at stations 4 to 11 (Figure 3.1) during the afternoon on July 6, 10, 13, 16, 19, 22 and 24. Maximum Wetness Index (MWI) is also shown and is represents the largest of the west or east bank TWI value for each longitudinal position.
Appendix D
Isotope data

This appendix section includes raw isotopic values for all sampled days for the West River and associated catchment between morning and afternoon.

Table D-1: Summary of δ¹⁸O and δD sample data for the afternoon of July 6, 10, 13, 16, 19, 22 and 24 between stations 4 to 11 and tributary inflows.

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Table D-2: Summary of δ¹⁸O and δD sample data for the afternoon of July 6, 10, 13, 16, 19, 22 and 24 between stations 4 to 11 and tributary inflows.

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Appendix E

Wetness indexes and inflow distribution

This section is composed of the remaining TWI figures not included in Chapter 3. The figures include a comparison of TWI and inflow types, and a comparison of MWI and inflow types.

Figure E-1: Topographic wetness index (TWI) for the study reaches of the West River for the east and west banks. Also shown is ΔT and inferred source inflows on the morning of July 19th. The contributing bank of surface and subsurface inflows is inferred on TWI levels. Note the inverted scale for the east bank TWI data.
Figure E-2: Temperature change (ΔT) and maximum wetness index compared in a regression tree analysis for the afternoon of July 19 inferred inflows in the study reaches. The total range of surface, subsurface inflows and channel snow are delineated by boxes, forming three main categories.
Figure E-3: Temperature change ($\Delta T$) and maximum wetness index compared in a regression tree analysis for the rain event of July 16 inferred inflows in the study reaches. The total range of surface, subsurface inflows and channel snow are delineated by boxes, forming three main categories.
Figure E-4: Temperature change (ΔT) and maximum wetness index compared in a regression tree analysis for the afternoon of July 10 inferred inflows in the study reaches. The total range of surface, subsurface inflows and channel snow are delineated by boxes, forming three main categories.
Figure E-5: Temperature change (ΔT) and maximum wetness index compared in a regression tree analysis for the morning of July 10 inferred inflows in the study reaches. The total range of surface, subsurface inflows and channel snow are delineated by boxes, forming three main categories.