The seasonal and inter-annual variation in water export from a boreal forest headwater stream

by

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ABSTRACT

Headwater streams are important locations representing access points to assess aquatic to terrestrial fluxes and source areas that control the water and solute fluxes to downstream aquatic environments. Boreal headwater streams are in post-glacial high latitude landscapes which show a wide range of topographical heterogeneity in landscapes containing high stores of organic matter highly susceptible to climate change. Boreal streams represent important sentinels for landscape changes relevant to global climate feedbacks. Catchment hydrology of boreal streams is controlled by the precipitation regime, particularly snow dynamics and the connection between the landscape elements and hydrology. Snowfall and snowpack dynamics are important features impacted by recent climate change over the last recent decades with potential to significantly impact catchment hydrology of boreal streams and thereby terrestrial to aquatic transport of water, solute, and nutrients in these high latitude landscapes. In this study, I investigated two key boreal headwater components (low gradient wetland and pond and upland forest regions) making up one third order stream watershed situated in western Newfoundland and Labrador in Canada. The second chapter describes a series of approaches evaluated for extending the rating curves required to obtain discharge estimates from continuously monitoring water level in remote, rocky small boreal streams beyond field measurement limited by access during key high flow periods. Results of this study suggest that the use of Ferguson flow resistant equations, which help to better inform the discharge-gauge height relationship through incorporation of stream geometry including

stream cross-section, stream gradient, and stream bed grain size data, significantly improves discharge estimates. The third chapter focuses on the study of water flux dynamics of a third order stream in three years contrasting in winter and snowmelt dynamics. The goal of this work was to assess how annual and seasonal discharge and runoff ratios of this small stream varied with (1) differences in snowpack dynamics (i.e. little winter melt to complete snowpack loss in mid-winter); and (2) watershed attributes defined by two key boreal watershed elements (i.e. low gradient pond and wetland dominated and hillslope forest dominated regions). Results from this work indicate that reductions in spring snowmelt events can lead to reduced discharge on an annual basis and decrease the contribution of water from upland hillslope regions that rely on snowpack water storage in this catchment. This suggests reduction in hillslope lateral flow water inputs associated with a reduced spring snowmelt event may be associated with (1) reductions in overall stream water discharge, and (2) increases in the proportion of low gradient pond or wetland contributions to the aquatic environment in these boreal landscapes. Both of these may have important implications for the boreal terrestrial-aquatic fluxes as well as downstream aquatic ecosystems in this region in a future warmer climate.

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List of Abbreviations and Symbols

N - North

W - West

km - kilometers

km^2 – Square kilo meters

USA - United States of America

mm - millimeters

mm^3 - cubic millimeters

cm - centimeters

m - metes

m^2 - square meters

$m\ s^{-1}$ - meters per second

$m^3\ s^{-1}$ - cubic meters per second

C - Carbon

g - grams

$g\ C\ m^{-2}$ - grams carbon per square meter

Pg - peta grams

CO_2 - carbon dioxide

DOC - Dissolved Organic Carbon

^{0}C - celsius

µm - micrometers

HPLC - High Performance Liquid Chromatography

TOC - Total Organic Carbon

Chapter 1.0 Introduction and overview

Despite the varying definitions of what constitutes a headwater stream, these streams are mainly identified as all first to second order streams which aggregate to compose two-thirds of the total stream length in a river network (Jencso et al., 2009). Therefore, headwater streams are the main contributors of sediments, and nutrients for larger watersheds by interacting with many hydrological pathways and source areas that control the downstream water and solute fluxes (Sidle *et al.,* 2000; Monteith *et al.,* 2007; Sorensen *et al.,* 2009; Huang *et al.,* 2013). Headwater streams are important locations for hydrologic connectivity within watersheds, but they are also very responsive to environmental changes occurring in their watersheds (Lowe and Likens, 2005; Reichstein *et al.*, 2013; Singh *et al.*, 2016). The responses of these streams, in terms of water and solute fluxes, to such changes, is critical to downstream ecosystems but can also tell us quite a lot about the surrounding landscape response as well. To better understand solute flux responses, such as carbon and nutrient transport, we first need to understand the hydrological responses of these important components of the watershed.

Boreal streams are in post-glacial landscapes that contain a lot of topographical heterogeneity and are responding to rapid climate change. The catchment hydrology of headwater streams is largely controlled by the precipitation regime (Mcglynn et al., 2004), surface and subsurface soil characteristics, the connection of landscape elements to topography (McGlynn and McDonnell, 2003; Seibert and Mcglynn, 2007), and hydrochemical attributes (Jencso et al., 2009). However, the connection between the catchment structure, precipitation regime and the runoff response are still poorly understood. This is partly due to a lack of discharge measurements for small streams particularly in more remote boreal regions. The majority of discharge stations set up by governments are focused on larger stream and rivers as they cover greater watershed areas and are generally simpler in some respects to monitor (e.g. accessibility downstream, rating curve development).

The hydrologic connection between the landscape elements such as hillslope, riparian zones, and streams occur when the water table continuity occurs between their interfaces (Mcglynn et al., 2004; McGlynn and McDonnell, 2003). The surface gradient associated with low relief and steep hillslope components of the landscape likely represents a significant control on the infiltration and runoff regime (Sivapalan et al., 1987; McDonnell et al., 2007). As an example, due to high flow conditions, hillslope soils can be more transitive and large quantities of water and solutes can be transported to downstream regions within those conditions. This connection is associated with catchment scale runoff production (Devito *et al.,* 1996), and the mechanisms for the delivery of water, sediments, dissolved organic matter and nutrients to streams (Creed *et al.*, 1996; McGlynn et al., 2002; McGlynn and McDonnell, 2003). Hillslope contributes as a major landscape element in forested headwater streams and hillslope soil acts as an important factor of transporting surface and subsurface water and solute fluxes to nearby streams (McGlynn and McDonnell, 2003; Jencso *et al.,* 2009).

Snowpack in boreal high latitude ecosystems plays a key role in the hydrological cycle where water is stored in the winter and released with the spring snowmelt (Mote et al., 2005; Schelker et al., 2013). Winter and spring melt discharge act as major components of the hydrology in snow dominated watersheds. The rapid response to winter and spring melt events influences the catchment hydrology of these snow dominated catchments and represents an important example of precipitation regime controls on hydrology (McGlynn and McDonnell, 2003; Schelker et al., 2013). Soil type and the soil water storage also regulate the runoff generation and transport of water to streams (Tsuboyama et al., 1994). Macropore flow within the forested soils can activate preferential flow paths especially during the high flow conditions that can enhance the delivery of water from terrestrial to aquatic environments (Jencso et al., 2009; Tsuboyama et al., 1994; Tsukamoto and Ohta, 1988).

Timing and the form of the winter precipitation have changed due to widespread warming in North America in twentieth century (Folland et al., 2001). Snow accumulation and snowmelt conditions are climatically sensitive features and therefore, changes in snow accumulation and snowmelt conditions make significant effects for the timing and the magnitude of the spring melt driven runoff (Nijssen et al., 2001; Stewart et al., 2005). Terrestrial to aquatic transport of water, solute, and nutrient fluxes are greatly influenced by the observed changes in snowmelt trends (Worrall et al., 2004; Kohler et al., 2009). For example, variability in DOC fluxes in streams and rivers documented in many studies as a consequence suggest the changing climate in high latitude ecosystems will likely have an impact on stream DOC export (Raymond and Saiers, 2010; Lepisto *et al.,* 2014). As an example, Huntington (2016) reported that increased runoff in winter and in association with increased winter rainfall is responsible for increased winter DOC fluxes and consequentially annual DOC fluxes of watersheds in one high latitude temperate region (Huntington *et al.,* 2016). The relative importance of those factors controlling the water and solute fluxes of high latitude streams is poorly defined but required if we are to determine responses of stream fluxes to climate change. Difficulties in obtaining accurate stream discharge for full range of gauge heights, however, remains a key challenge in defining those fluxes and monitoring their responses to such change.

Headwater streams are often high gradient steep streams with large bed particles and measuring direct discharge at high flow conditions has always been a challenge due to the inaccessibility of field sites particularly during conditions of spring melt and major storm events (Lindner and Miller, 2012; Sivapragasam and Muttil, 2005). The development of appropriate methods of predicting discharge for unmeasured gauge heights at high flow conditions will lead to accurate quantification of headwater stream runoff to investigate the link between the headwater streams and catchment hydrology. Accurate quantification of headwater runoff can be important to understand the behaviour and effects of landscape elements in catchment hydrology (Sivapragasam and Muttil, 2005; Shiklomanov *et al.*, 2006; Lindner and Miller, 2012; Coz *et al.*, 2014). Continuous records of gauge height data can be transformed into continuous discharge data using an accurate direct stage-discharge relationship in order to obtain reliable quantifications of stream discharge and runoff (Ramsbottom and Whitlow, 2003; Lang *et al.*, 2010).

Boreal forests are high latitude ecosystems where some of the most significant effects of climate change are predicted to occur (Hansen et al., 2006; Hassol et al., 2004; Serreze et al., 2000; Smithson, 2002). One-third of the area of the boreal zone is located in Canada and Alaska and its unique cold temperature conditions create better conditions for the storage of one of the largest reservoirs of global terrestrial carbon (30-35%) as SOC stocks (Ruckstuhl *et al.,* 2008). Some climate predictions based on future climate scenarios mentioned that the largest temperature and precipitation increases will be in northern hemisphere upper-latitude regions (IPCC, 2001; Smithson, 2002). Changes in carbon fluxes have been measured in boreal forested watershed regions in the past few decades, which are interpreted to be in response to the climate induced changes (Stocks *et al.*, 2000; Ruckstuhl et al., 2008).

Terrestrial DOC is a major form of organic matter transported through streams, especially in headwater streams due to their interaction between the surrounding terrestrial environments. Stream DOC flux and its temporal and spatial variations are governed by stream discharge and soil organic carbon (SOC) characteristics. For example, riparian SOC content can be a significant factor that controls 90% of DOC flux in streams as it has a significant effect on DOC fluxes in some streams (Dosskey and Bertsch, 1994; Huang *et al.,* 2013). In addition, soil temperature, soil moisture, atmospheric nitrogen and sulphate deposition, plants inputs into the soil and vegetation cover can also affect stream DOC composition and flux (Eimers *et al.,* 2008). Precipitation affects DOC concentrations both directly and indirectly. In peaty soils, carbon is available and high rainfall rates put DOC into solution. Thus, the highest DOC concentrations often coincide with high discharge values associated with higher precipitation (Laudon *et al.,* 2004; Worrall *et al.,* 2006). Soil saturation and high water tables can activate surface and subsurface flow paths that may efficiently connect new soil carbon sources to streams, thus increasing DOC concentrations in many high flow events (Fiebig et al., 1990; Worrall *et al.,* 2006; Jennings *et al.*, 2010). Furthermore, these activated flow paths may influence the stream chemistry by governing the fluid residence times and the chemical equilibrium of stream water within the catchment.

DOC concentration and export in surface water have been increasing over the past few decades in many high latitude regions of North America and Europe (Hongve et al., 2004; Evans et al., 2005; Monteith et al., 2007). As a result, multiple mechanisms for this widespread increase in DOC concentration have been proposed. Proposed mechanisms include: increasing temperature and higher atmospheric CO_2 (Freeman *et al.*, 2001), recovery from acid deposition (Monteith et al., 2007), changes in sea-salt deposition (Freeman et al., 2001), alteration of precipitation regime (Freeman et al., 2001; Hongve et al., 2004), and combination of these factors have been suggested as possible leading factors for increasing DOC concentrations in streams (Hongve et al., 2004). Changes in either or both temperature and discharge may cause short-term temporal variability in DOC in running waters (Clark et al., 2005; Eimers et al., 2008; Kohler et al., 2009). Many authors suggested direct and indirect effects of climate drivers impacting the flux or concentration of DOC in higher latitude watersheds. Shifts in the precipitation regime and changes in snowpack and snowmelt dynamics control discharge and runoff ratios and thereby, can have impacts on the flow paths, source and processing of solutes en route to the aquatic environment. These factors suggest the potential for changes in solute transport including DOC as a consequence of the changes in the snowpack.

Newfoundland and Labrador boreal forest watersheds enable documentation of the water and solutes in boreal landscapes and effects of temperature and precipitation changes on these ecosystems at watershed scales. The climate in this area is characterized with cold winters covered with snow and the mean annual precipitation is approximately 1200 mm with an average annual daily temperature of 4.0 ^{0}C. Pynn's Brook Experimental Watershed is located in the mesic NW Atlantic boreal region of Canada which allows investigation of stream discharge dynamics of a headwater boreal stream within a mesic climate. Furthermore, this study site enables the study of these water fluxes specific to two important boreal landscape elements and also provide tools to investigate water fluxes within the context of more specific hillslope hydrologic studies occurring within the same experimental watershed (e.g. soil and ground water and solute fluxes).

The overall objective of this thesis is to quantify the annual and interannual variation of water and the DOC fluxes in one boreal headwater stream. By quantifying the variation of headwater stream water and DOC fluxes, I addressed the seasonal trends of these fluxes associated with estimated water fluxes and how these trends differ between two key landscape elements. Water and DOC fluxes were evaluated using the regional meteorological variables (precipitation, runoff ratios) to understand the variation of these fluxes with respect to regional scale. These variations provide insight into how some climate change mechanisms, such as rain or snowmelt in winter, an increase in growing season length, or the timing of snowmelt, might affect soil and stream water and DOC fluxes.

The main objective of the second chapter was to establish a suitable method for predicting discharge measurements across the full range of observed gauge height values to obtain an accurate stage-discharge relationship for two sites on a rocky headwater stream. I compared the predicted discharge values with the regional runoff ratios to assess the estimated discharge values and rating curves were further assessed by determining the full range of the discharge values considering the factors used to generate the methods. By applying the rating curve procedure developed in Chapter 2, I investigated the variation in water flux of a boreal forested watershed in three years experiencing contrasting winter precipitation regimes, and specifically within two common but contrasting components of the watershed. Furthermore, I focused on how seasons and precipitation regimes impacted water flux contributions from different landscape components of the catchment.

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Chapter 2. 0 Development of an approach for obtaining rating curves for rocky boreal headwater streams

Abstract

Boreal headwater streams are steep with large bed particles or bedrock which make the measurements and estimation of the discharge difficult. Rating curves are often used and extended (extrapolated) beyond the highest gauge heights measured at the field sites due to the inaccessibility to field sites during the floods and higher flows. Such extrapolations, commonly used in large rivers, with more homogenous beds and geometries, are not likely to accurately represent gauge height to flow rate relationships at high gauge heights in small rocky streams. In this study, I tested several extrapolation techniques such as curve fitting methods using simple regression techniques, Manning's n method, Jarrett's method, a combination of Manning's and Jarrett's method, and Ferguson resistance equation to extend the rating curves beyond the field discharge measurements. Results of this study suggest the Ferguson resistance equation for extrapolating rating curves based on Ferguson flow resistant provides the most accurate rating curve to apply to the small rocky headwater stream studied here. Obtained discharge estimates from the Ferguson resistance equation were further verified by calculating runoff ratios and comparing those ratios with regional runoff ratios. The result obtained here is likely due to the fact that the flow resistant approach further constrains the rating curve more through site specific field data such as stream cross-section, stream gradient, and grain size data.

2.1. Introduction

Headwater streams link terrestrial and aquatic ecosystems and thus are important locations representing the interface between water bodies and the terrestrial environment (Devito et al., 2005; Gomi et al., 2006; King et al., 2010). Defined as first or second order streams, headwater streams compromise two thirds of the total stream length worldwide and the total watershed area (Leopold et al., 1964; Sidle et al., 2000). What traverses this interface and ultimately observed within the headwaters is highly sensitive to changes in hydrology and biogeochemical processes occurring within the watershed (Fiebig et al., 1990; Clark *et al.*, 2005; Lowe and Likens, 2005; Eimers *et al.,* 2008; Futter and Wit, 2008; Reichstein *et al.*, 2013; Singh *et al.*, 2016). Therefore, the behavior of solute fluxes within headwater streams can assist us in understanding terrestrial to aquatic connectivity and what controls this exchange (King et al., 2009).

Recent studies suggest that significant lateral and vertical fluxes of carbon occur within the headwaters of some boreal and arctic watersheds (Köhler et al., 2008). Changes in processes of lateral and vertical fluxes may cause increases or decreases in watershed export of carbon. The downstream reaches of some high latitude rivers are exhibiting increased organic carbon fluxes, however, the factors contributing to this increased “browning” remains debated (Jennings et al., 2010). Increasing trends, especially for carbon fluxes have been reported by many authors and potential drivers of these trends include rainfall, temperature, acid deposition, land-use and enrichment of CO_2. Some rivers, such as the Yukon river exhibit decreasing carbon fluxes (Déry and Wood, 2005). This spatial variability suggests that these changes in carbon fluxes in high latitude regions may be primarily driven by precipitation and evapotranspiration changes. In some instances, these increases are attributed to increased transport in winter suggesting potential climate shifts in hydrology associated with changing winter conditions (e.g. trends of decreased snow and increased rain) (Huntington *et al.*, 2004; McCabe *et al.,* 2018). Winter rainfall can lead to snowpack reduction and a greater proportion of organic matter can be mobilized before the typical snowmelt period, this may increase terrestrial to aquatic carbon fluxes (Eimers *et al.,* 2008; Berghuijs *et al.,* 2014).

Watershed morphology and soil permeability significantly regulate hydrological processes, thereby controlling discharge patterns (Jencso et al., 2009; Schelker et al., 2013). Headwater streams in post glaciated watersheds often drain from pond, lakes and wetlands, before traversing through gradient hillslopes. These watersheds exhibit a wide range of landscape heterogeneity leading to complexity in watershed morphology that, combined with variations in vegetation, bedrock, and soil type govern the catchment hydrology in these recently deglaciated landscapes (Devito *et al.,* 2005; Freeman *et al.,* 2007; Mcdonnell *et al.*, 2007; Schelker *et al.*, 2013). For example, reduced soil water storage capacity in impermeable layers of frozen soil in boreal forest catchments leads to preferential paths ways to the generation of runoff (Carey and Woo, 1999; Devito et al., 2005) and thereby related solute sources and fluxes.

Understanding the interactive roles of such features, including soil water storage and catchment morphology, with precipitation patterns can aid in our predictive understanding of runoff generation in headwater streams in particular as a response to climate change. Snowpack formation and melt typically dominate many high-latitude watersheds with spring melt events often representing >70% of total runoff (Hjalmar Laudon et al., 2004). Variation in snowfall and snowpack associated with climate change represents an important change in precipitation and thereby runoff patterns. Monitoring headwater runoff patterns in association with variation in precipitation dynamics can be important to a predictive understanding of runoff generation with climate change.

Boreal headwater streams are often steep with large bed particles or bedrock, this makes the measurement and estimation of the velocity and thus discharge difficult. Reliable stage data can be obtained from continuously monitored water stage, but the transformation of these data into continuous discharge requires an accurate stage-discharge relationship (McMillan *et al.,* 2012). Field measurements of velocity and calculated discharge from velocity and area measurements are correlated with observed gauge heights to obtain the stage-discharge relationship, which is referred as the rating curve (Lang et al., 2010).

Accurate rating curves require a stable channel cross section and a range of empirical discharge values throughout the range of observed gauge heights (Le Coz et al., 2014; Lindner and Miller, 2012; Rantz, 1982; Shiklomanov et al., 2006; Sivapragasam and Muttil, 2005). Discharge measurements during the high flow levels may be difficult due to dangerous conditions and the inaccessibility of field sites (Sivapragasam and Muttil, 2005); conditions typical of headwater streams during snowmelt or major storms in boreal landscapes. Moreover, stream gauging structures (e.g. weirs) are difficult to build and maintained in remote regions that are susceptible to ice damage and floods (Lang et al., 2010; Lindner and Miller, 2012). Therefore, empirical field discharge values for high stages are often not obtained, and indirect estimates are required. Therefore, suitable techniques for estimating discharge values for gauge heights above the measured field discharge values need to be developed to study boreal headwater stream runoff (Braca, 2008; McMillan et al., 2010; Ramsbottom and Whitlow, 2003; Shiklomanov et al., 2006; Sikorska et al., 2013).

Estimation of discharge values for unmeasured gauges in large river systems is quite common and has been applied using a few different indirect methods (Franchini *et al.,* 1999; Toronto and Region Conservation Authority, 2008; Medalie, 2013; Roy and Mistri, 2013) and such methods include simple regression analysis, and Manning's equation-based methods. Significantly different approaches were used in Manning's based methods rather than regression methods as unmeasured velocities are estimated using flow resistance equations which are then used with the measured cross-sectional area to compute discharge. Graphical methods based on logarithmic or regression methods such as power regression methods are a common and simple way of establishing the stage-discharge relationship (Herschy, 1995, 1993). DeGagne et al. (1996) used a linear regression method to linearized stage-discharge data which is primarily used in rating curve extending. As an example, among the various curve fitting methods of extrapolation, polynomial regression of second and third degree is often used (Torsten et al., 2002).

In previous studies, methods that constrain flow resistance equations (e.g. the Manning's equation) were more accurate than regression models in predicting high discharges. Flow resistance methods, however, require additional information about stream channel roughness and gradient (Arcement and Schneider, 1983; Roberts et al., 2016). Headwater streams are high gradient streams and applying flow resistance equations to rocky headwater streams is quite challenging due to the need of appropriate Manning's roughness coefficient values (n) and need of accurate measurements of slope changes. Several approaches were made by Jarrett (1987) and Ferguson (2006) to develop flow resistance equations. Jarrett (1987) defined Manning's n values for high gradient streams based on empirical data and this relationship was developed between Manning's n and friction slope and hydraulic radius. Jarrett (1987) stated that the roughness coefficient is a measure of the factors that effects on flow resistance such as bed material, channel geometry, and vegetation type and density. Ferguson (2007) further developed flow resistant equations based on grain size data. More filed parameters such as grain size of the stream, hydraulic radius, and channel gradient were used in his approach, providing more accuracy to velocity estimates for high gradient streams. The flow resistance was related to relative roughness (d/D_{84}) and calculated by average depth (d) divided by D_{84} values for boulder bed streams as recommended by Ferguson (2007).

The main objective of this research was to establish a suitable method for predicting discharge measurements across the full range of observed gauge height values to obtain stage-discharge relationships for two sites on a rocky headwater stream. To accomplish this, I used both extrapolations (curve fitting), and flow resistance methods. I also integrated the two approaches by estimating high flow values with the flow resistance approaches. This resulted in five different extrapolated stage-discharge relationships using a curve fitting, flow conveyance using Manning's equation (FC method), velocity estimates based on Jarrett's n equation (Jarrett's n Method), the combination of FC and Jarrett's method (combined method), and finally Ferguson's resistance equation (Ferguson's method). In order to obtain five relationships, I used grain size, channel cross section, and channel gradient data to compute the velocities and then discharge values. These five curves were generated for one site dominated by forested hillslope watershed and an upper site dominated by low relief pond dominated watershed**.** Resultant rating curves were assessed by three main steps. Initially, predicted discharge values were simply compared to one another using the existing knowledge of the rating curves to narrow down the best methods. Furthermore, the precision of each rating curve was used to bring out the best approaches. Best described curves were then assessed by comparing them with the regional runoff values. I then applied the best rating curve to three separate years of continuous gauge height data to calculate water fluxes and compare with the runoff in each year. Comparisons to the regional runoff ratios enabled me to further assess the validity of the approaches developed.

2.2 Methodology

2.2.1. Study region and site description

Horseshoe Brook is part of the Pynn's Brook Experimental Watershed (having coordinates of 48° 53' 14" N, 63° 24' 24" W) site a collaborative research platform supported by the Canadian Forest Service and part of the Newfoundland and Labrador Boreal Ecosystem Latitudinal Transect (NL-BELT). Pond dominated upper Horseshoe Brook site (UHS) (49° 4'' 8.4" N, 57° 20' 58" W), and forest hillslope dominated lower Horseshoe Brook site (LHS) (49° 4' 9.6" N, 57° 20' 55" W) were the main focus of this study. Catchment areas for UHS and LHS sites are similar (3.84 km^2 and 7.52 km^2, respectively), but differ in catchment characteristics such as morphology. UHS is a low gradient region influenced by wetlands and five ponds whereas the LHS catchment is dominated by forested hillslopes that connect to an incised, boulder-bed stream channel without a prominent floodplain. The study site is dominated by Black spruce forest (*Picea maiana*) varying from mature (>80 years of age) to recently harvested within 10 years (Moroni et al., 2010; Bowering et al., 2019).

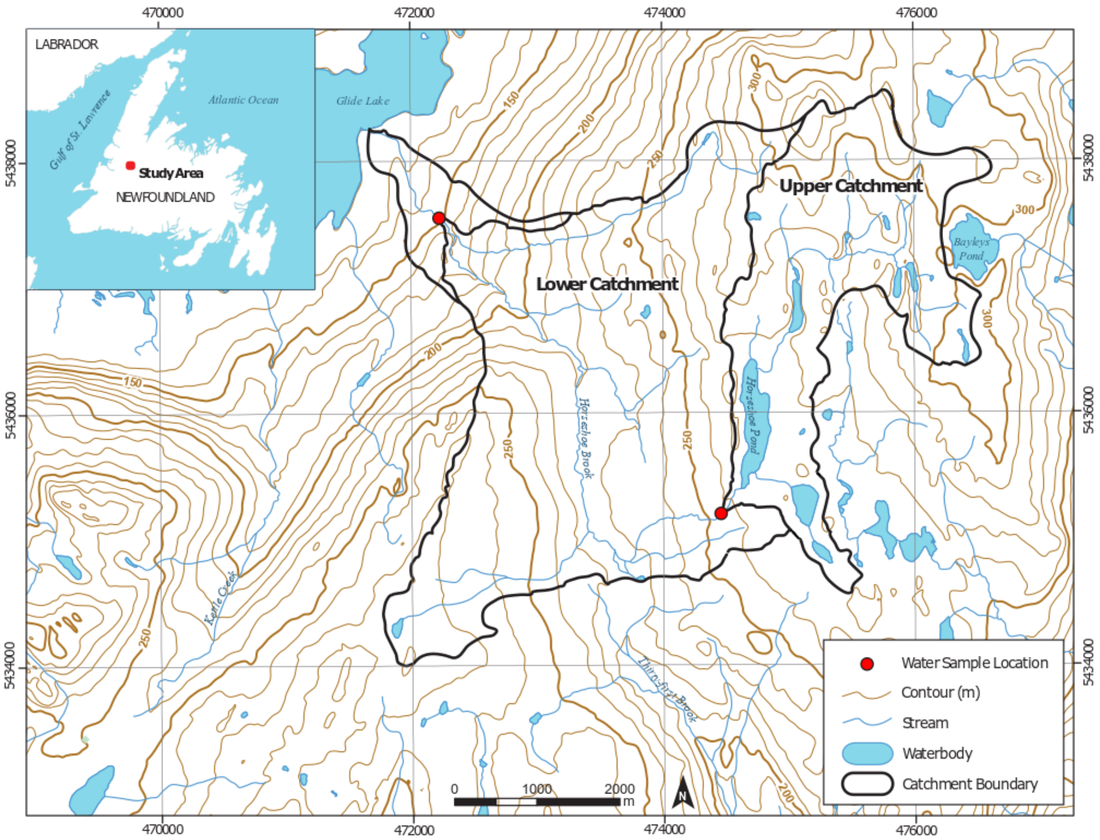


Figure 1. A map of the catchment area for the Horseshoe Brook located at the Pynn's Brook experimental watershed area near Deer lake, western Newfoundland and Labrador, Canada. (Upper and the lower portion were labeled as UHS and hillslope, respectively.

2.2.2 Stream site data collection

Continuous water level data on stream temperature and water level were collected over three separate years between October 2014 to October 2018 using in situ pressure transducer dataloggers (Onset U20 water level logger; MA, USA). Four field site visits per year to stream sites were conducted for field measurements of discharge, and cross-section, the slope, and roughness of the channel and to download data during this period. Friction slope or the slope of the channel was also obtained from the onsite surveys where the cross-section of the stream was located. Stream cross section information was made two times during 2017 and 2018 to assess the possible changes of the stream channel. Stream gradient measurements were also taken by following the auto levelling surveying techniques at each stream site.

Stream discharge was measured at the field according to the velocity-area method (Turnipseed and Sauer, 2010). The velocity of the flowing water was measured using a current meter (Swoffer model, 3000, Washington, USA). Width of the stream was divided into 10-15 number of increments depending on the depth and the velocity of the stream. Stream depth, and average velocity were measured for each increment. For streambed where the depths are shallow, velocity was measured at 60% of the distance from the water surface to the stream bed. Grain size data were collected using a simple method suggested by Wolman (1954). Total roughness of the channel can be expressed as relative roughness (d/D_{84}) of the channel (Leopold et al., 1984), where d is average depth and D_{84} is particle intermediate diameter which equals or exceeds 84 percent of particles. Larger grains contribute to more grain resistance and therefore, the flow resistant was largely determined by the grain size of the stream bed (Prestegaard, 1983). The grain size of the surface material was measured by walking a grid pattern over the site. Particles under the toe of the boot were picked up and the intermediate axis of the particles was measured. Measurements were taken approximately from around 75- 100 particles and data were plotted to compute the D_{84} particle intermediate diameter.

Continuous water level records were captured for 30 min time interval at the field by the HOBO water level loggers that were installed at the stream sites. Downloaded data from the probes were atmospheric corrected using the atmospheric pressure data downloaded from atmospheric pressure loggers located at each stream site using HOBO air pressure data logger (Onset RX3000 data logger, MA, USA). However, there were no atmospheric probes were installed at the sites in 2014 therefore, the source for atmospheric correction for 2014 was downloaded atmospheric data from the Deer lake station (station number-02YL007) Environmental Canada website.

The three years of gauge height and discharge data between 2014 – 2018 were studied representing three different hydrological years based on the variation of the precipitation (Table 1). Year-1 was a hydrological year with a consistent and continuous snowpack and large spring snowmelt without any real melting events during the winter. Year-2 and year-3 were significantly different from year-1 with different winters having a couple of storms or melting events during the winter period. In Year-2, 4 melting events occurred throughout the winter and in Year-3 a significantly reduced snowpack was experienced as a result of a catastrophic melting and rain event that occurred in mid-winter (January 10th - 30th). Amount of snow to the total precipitation was reduced from 42% in Year 1 to 34% in years 2 and 3.

Table 1. Meteorological data downloaded for three years (Year 1 - 2014 October to 2015 October, Year 2 – 2016 October to 2017 October, and Year 3 – 2017 October to 2018 October).

Period	Total annual precipitation (mm)	Number of snow melting events	Amount of snow to total precipitation
Year 1 (continuous snowpack)	1581.4	Spring melt only	42%
Year 2 (Small mid winter melts)	1285.7	4 mid winter + spring melt	34%
Year 3 (Large mid winter melt)	1504.7	One major mid winter + spring melt	34%

2.2.3. Calculating stream cross sectional area and wetted perimeter

Cross section areas for UHS and LHS horseshoe sites were calculated using the field observations for the gauge heights which begin from the bottom of the channel to the highest point where the cross-sectional area was recorded. The place where the probe was located was considered as the bottom of the cross-section which was considered as the zero-water level. Bank full depth was traced from the continuous water level records and considered as the maximum water level that was recorded in the probes installed at the field sites. Calculation of the wetted perimeter for each gauge height was conducted using the "Pythagoras theory," and the determined wetted perimeter for each section (L) of the whole cross section was finally added to the total wetted perimeter of each gauge height. Computed wetted perimeter for a gauge height was then checked by an equation which is used to calculate the wetted perimeter of a channel based on the shape of the channel (Eq. 2).

$$P = T + \frac{8y^2}{3T} \qquad (2)$$

Where, T is the distance from one end to the other end of the channel, y is the channel depth at the deepest point, z is the depth at specific water level, and x is the distance from the deepest point to the channel end.

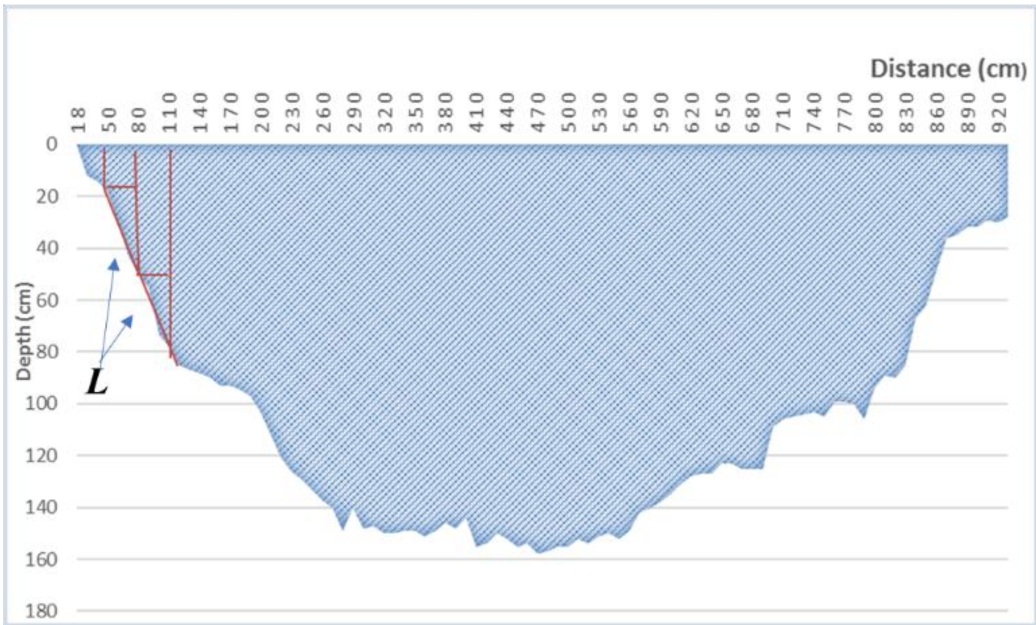


Figure 2. Illustration of the cross-sectional area approach and calculations of cross section of each small segment

2.2.4. Methods used in this study to extrapolate rating curves beyond the field measurements

Rating curves were developed for each stream sites. However, as noted earlier a narrow range of discharge measurements was captured at the field which covered a range of 10 cm to 49 cm of gauge heights for each site. Maximum possible gauge heights were 135 cm to 155 cm for UHS, and LHS respectively. Therefore, a need for a representative rating curve for the entire range of water levels often highlighted. Five approaches such as curve fitting, FC, Jarrett's, a combined method, and Ferguson's methods were tried to predict the unmeasured discharge measurements.

The following relationships were derived using gauge height, cross-sectional area and the water velocity for each site in order to apply and assess five approaches for extending the rating curve for these two sites in order to calculate the discharge.

I. A relationship between gauge height vs. inundated cross-sectional area, which provides information on the increase in cross-sectional area with gauge height (Q = A*V, where Q = discharge m^3 s^{-1}, A = cross sectional area m^2, and V = velocity m s^{-1}).

II. A relationship between gauge heights vs. velocity for each site.

This relationship was generated by calculating velocity using the measured discharge data (m^3 s^{-1}). Each discharge measurement was divided by the cross-sectional area (m^2) to determine the mean velocity.

Velocity usually increases as a power function of discharge for in-channel flows (Leopold et al., 1968; Dingman et al, 1984; Lawrence, 2007a) Therefore, I used geometry relationships to determine if either site might have sufficient data to evaluate this relationship between discharge and velocity (Leopold et al., 1968).

III. The two relationships, Q–area (A) and Q–velocity (V) were constructed for each site using a range of gauge height values that encompassed all the observations of Q made at each site. This resulted in two sets of equations:

$$V = kQ^{m} \tag{3}$$

$$A = oQ^{p.} \tag{4}$$

Due to continuity (i.e. Q = AV), the exponents of these two equations must sum to 1, and the product of the coefficients must equal to 1. Exponents and the coefficients were tested using five methods for each stream site and the values of the exponents and coefficients were evaluated to determine the continuity of the Q = A*V equation.

2.2.5. Methods used to extend the field discharge measurements to generate rating curves

Five methods have been applied in this study including one graphical method and four methods based on stream geometry information to assess their adequacy in generating discharge estimation in a small, rocky boreal stream. Among those methods, curve fitting relied on regression relationships and except for curve fitting method flow conveyance, Jarrett's n method, combined method and Ferguson's method were based on the data obtained from the geometry of the channel such as cross-sectional area, gradient, Manning's roughness coefficient and the average grain size.

2.2.6. Regression methods

Generated rating curves using the field measured discharge values were extended by applying the second and third order polynomial regression models in order to estimate higher discharge values (Torsten et al., 2002).

2.2.7. Application of Manning's equation for FC method for rating curve extending

Manning's roughness coefficient, *n* is used to describe the relative roughness of a channel which is in the form of manning's the general equation for open channel flows (Jarrett, 1985; Barnes, 1967).

$$V = \frac{1{\cdot}486}{n} R^{\frac{2}{3}} \cdot S_f^{\frac{1}{2}} \tag{5}$$

Where, V is the mean velocity of flow (in feet per second), R is the hydraulic radius (in feet), S_f is the slope of the energy grade line, and n is the Manning's roughness coefficient.

Manning's roughness coefficient was used here to predict the discharge values based upon hydraulic radius and the slope of the channel for each of the two sites. Hydraulic radius was computed by dividing the cross-sectional area of a stream from the wetted perimeter. It is important to keep in mind that the wetted perimeter does not include the surface irregularities of the submerged section. For sites where gradient and velocity are measured, n can be calculated.

For coarse-bed rivers, n usually decreases with discharge (Ferguson, 1986). Flow conveyance, *K* equals to $S^{1/2}/n$, can be calculated using Eq (6).

$$K = V \cdot R^{\frac{2}{3}} \qquad (6)$$

Where, *K* is flow conveyance, *V* is the velocity obtained from measured discharge, *R* hydraulic radius determined from the ratio of the cross-sectional area divided by the wetted perimeter. K values were determined using the Eq (6). Conveyance *K* for channels often increases with gauge height, but it can also stay constant if increases in flow turbulence offset the increase in flow depth. Flow conveyance was assessed to determine if it was relatively consistent or predictable for each site and thereby assess the potential of using K to predict unmeasured velocity values for gauge heights beyond the measured discharge values. The estimated *K* values for unmeasured gauge heights obtained via Equation 6 were multiplied by $R^{2/3}$ to obtain unmeasured velocities and then discharge values. Rating curves were then generated using those newly predicted discharge values based upon this approach constrained using Manning's equation and flow conveyance.

2.2.8. Application of Jarrett's n method for rating curve extrapolation

$$n = 0 \cdot 39 S_f^{0 \cdot 38} R^{-0 \cdot 16} \quad (7)$$

Where, n is roughness coefficient, S_f is the friction slope and R is the hydraulic radius.

Equation 7 was used for predicting the Manning's *n* for higher gradient natural channels such as rocky headwater streams which are graphically represented in Figure 3. The slope of the channel (S_f) was used in conjunction with *R* to obtain *n*. Resultant n value for each gauge height was used to estimate the velocity as per Eq. 5 and Eq. 7 and calculated *V* values were used with the channel cross-section to compute the discharge.

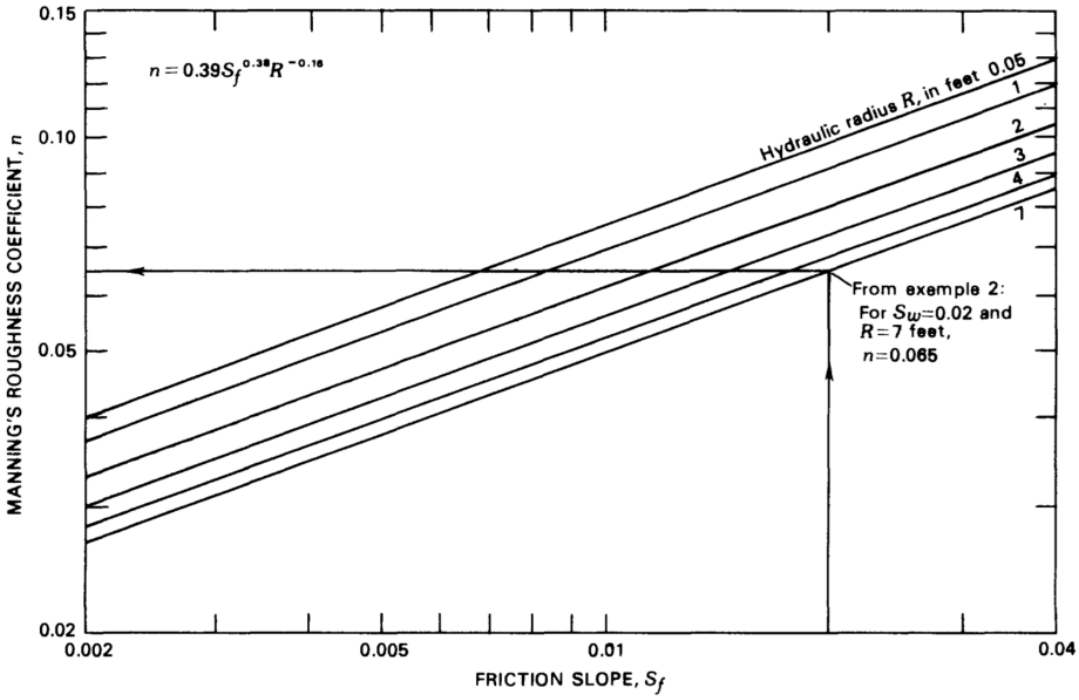


Figure 3. The figure of Manning's roughness coefficient vs. friction slope which is used to compute the Manning's n value taken directly from Jarrett (1987).

Resultant discharge values from FC and Jarrett’s methods were plotted against the field measurements of discharge values to check the validity and the deviation of the estimated values from the field discharge values. However, as reported in the literature, Jarrett’s method was most suitable for estimating high energy discharge values which were limited using this approach for estimating a range of discharge values (Jarrett, 1987).

2.5.4. Generating a combined rating curve using both FC and Jarrett’s n method

A combined rating curve was also generated using both sets of discharge values that were determined by the FC and Jarrett’s methods in order to obtain a reasonable representative rating curve for both lower and higher gauge heights. The lower part of the rating curve (up to field measurements) was generated using the discharge values obtained from the FC method, and the upper section (beyond the field measurements) was constructed with the discharge values obtained from Jarrett's method. The obtained gauge height-discharge relationship using a regression line through the resulting values from the two methods was adjusted by the solver function in Microsoft Excel to minimize the error of estimation. After generating a representative rating curve for both lower and higher discharge values, the curve was adjusted using “solver function” in Microsoft Excel to obtain the best fit curve for the stage-discharge relationship.

2.2.9. Application of Ferguson's resistance equation for rating curve extrapolation

Ferguson velocities and discharge values were estimated and used to generate another gauge height-discharge relationship. Those estimates were based upon flow resistance which was quantified and defined by relative roughness, cross section, average velocity, mean flow depth, and gradient (Ferguson, 2007). Hydraulic radius and grain size data or intermediate diameter of streambed particle distribution (D_{84} values) were used in this approach to computing the velocities and the discharge values. The flow resistance was related to relative roughness (d/D_{84}) and calculated by average depth (d) divided by D_{84} for boulder bed streams as described in Ferguson (2007).

$$u^* = g^{0.5} \cdot R^{0.5} \cdot S^{0.5} \tag{8}$$

Where u^* is shear stress, g is gravitational acceleration, R is hydraulic radius (cross section/wetted perimeter), and S is gradient. Hydraulic radius was determined by cross sectional area divided by wetted perimeter for each gauge height and shear velocity was calculated by multiplying the square root of g, R, and S (Eq. 8).

$$\frac{u}{u^*} = 2 \cdot \left(\frac{d}{D_{84}}\right) \tag{9}$$

Where u/u^* is a resistant parameter, d is depth, u is stream velocity and D_{84} is the particle intermediate diameter that equals or exceeds that of 84 percent of the particles. u/u^* values for each water levels were determined using Eq. (9) and resultant was multiplied by u^* values which were obtained from Eq. (8) to determine u. u value was then multiplied by the channel cross section to determine the discharge. In summary, each of the five separate methods described above utilized one or two different equations in order to predict discharge values associated with gauge heights at each of the two streams sites (Table 2).

Table 2. Equations used for different extrapolation methods

Method	Mathematical equation used
Curve fitting	Based on a regression method
FC method (Ferguson, 1986)	Eq. 6
Jarrett's method (Jarrett, 1987)	Eq. 7
Combined method	Eq. 6, Eq. 7
Ferguson method (Ferguson, 2007)	Eq. 8, Eq. 9

Table 3. Field data used for different extrapolation methods (all required gauge height and the available field discharge measurements).

Method	Required field data
Curve fitting	Gauge height and measure discharge values
FC method	Hydraulic radius
Jarrett's method	Hydraulic radius, channel gradient
Combined method	Hydraulic radius, channel gradient
Ferguson method	Hydraulic radius, channel gradient, grain size data

2.2.10 Assessment of rating curve generation

Five approaches were assessed in order to select the best rating curve which provides the most reliable estimates of unmeasured gauge heights. To accomplish this task, mainly three approaches were considered. Initially, predicted discharge values were simply compared with one another to rule out the inappropriate unrealistic discharge estimates. Next, rating curves were narrowed down calculating the precision of discharge estimates as the next step of assessing the curves. Lastly, discharge data were used to calculate runoff (Water Volume/Basin Area) and to compare with precipitation. This provided a water balance constraint; summer storm and annual runoff values could not be higher than precipitation values. Runoff ratios (runoff coefficients) could also be compared with regional values.

Runoff ratios were used to assess the suitability of each of the five methods tested here for extrapolating discharge values. Estimated discharge values were converted to mm of runoff in order to directly compare those estimates with the precipitation obtained from the Deer Lake Station, located 20 km from the study catchment, and operated by Environment Canada (02YL007, Environmental Canada). Obtained discharge values were then divided by the precipitation values to calculate the seasonal and annual runoff ratios. The annual runoff values for each site was compared with the regional runoff ratios computed from the Humber River, a 5th order river with a catchment area of 2,110 km^2 (Humber Village Station, 2YL001, Environment Canada 2010-2018). The study catchment is located within the Humber River watershed and 30 km from the Humber Village discharge station. Runoff ratios were also obtained for a local 3rd order stream, South Brook, as another reference to determine the validity of the calculated runoff values from each approach assessed here. The South Brook discharge station (02YL004, Environment Canada) is located approximately 35 km from the study catchment and captures the runoff from a catchment area of 58.5 km^2. However, computed runoff ratios were used to assess and select the best rating curve that provides the most reliable estimates.

2.2.11 Error analysis

Lower and upper ranges of discharge estimates were calculated for the best rating curve after accessing the five curves. Rating curve uncertainties depend on the data that were used to estimate discharge values for each method. Each of the factors that were used for generating rating curves has some level of uncertainty and the best estimate of uncertainty was to determine the uncertainty of each of the factors that were used to generate the rating curve. Although I am unable to determine the full uncertainty associated with the rating curves developed for each stream site, I determined the variation observed in each of the factors used and calculated the full range of known variation for the discharge estimates made.

I used the most recent field measurement of each parameter to develop rating curves as I had few measurements for each field parameter that were collected at different times. I determined the percentage variation of each of the parameter by considering the number of measurements obtained at the field. As an illustration, if a parameter was measured two times at the field, I used to the most recent value for rating curve development and then and the percentage variation of the most recent value to the previous measurement was calculated. After calculating the percentage variation, I determined the highest and the lowest possible error values for each parameter using the above percentage to obtain the highest and the lowest variation ranges.

2.3 Results and Discussion

2.3.1. Stream site characteristics and discharge

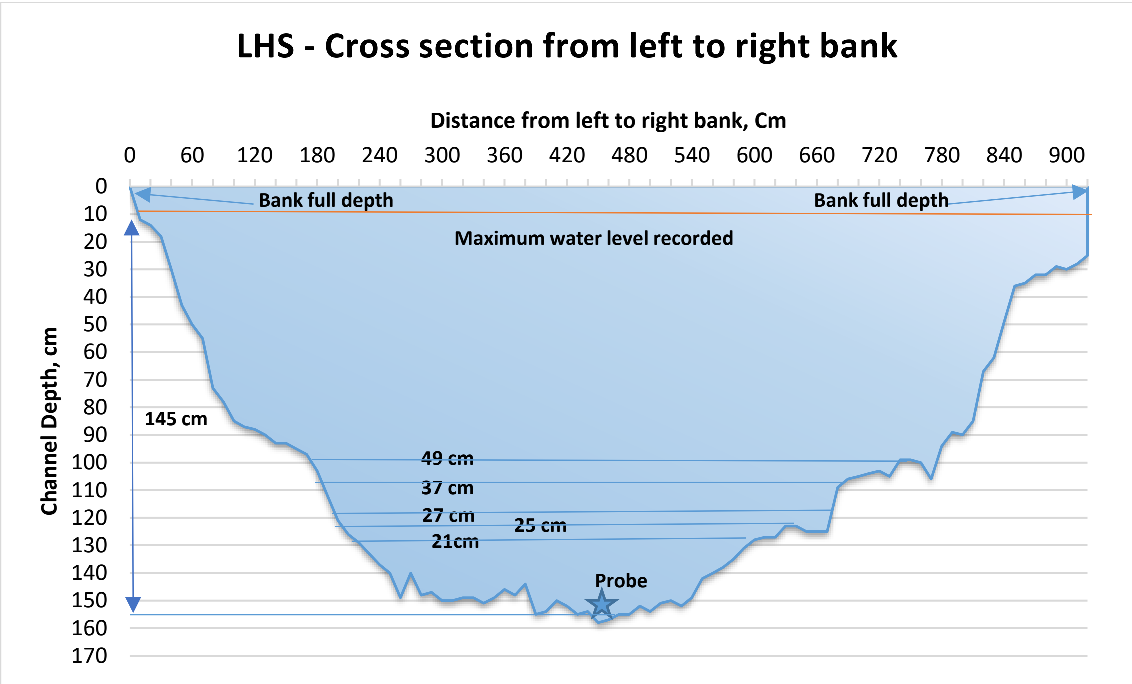
The UHS and LHS field sites differ in the shape of the cross section, stream gradient, and particle size distribution. The UHS site drains a low relief area with shallow first order streams consisting of rocky and boulder characteristics. Compared to the UHS stream site, LHS stream site drains primarily a hillslope region which helps to generate more runoff.

Bank full levels based upon cross sections were referred as gauge heights that just begins to spill out of the channel into flood plain, and those values for all three years were 145 cm and 85 cm, respectively, for LHS and UHS. Highest gauge heights for LHS and UHS were recorded in the probes as 139 cm and 72 cm, respectively and these values were below the bank full levels for both LHS and UHS stream sites (Table 4). Field discharge measurements were made for 9 and 5 different gauge heights at LHS and UHS, respectively. Field discharge measurements covered gauge heights between 21 and 49 cm at the LHS stream site and from between 10 cm to 15 cm at UHS. The LHS exhibited a comparatively higher gradient (0.031 m per m) relative to the UHS stream site (0.023 m per m). Intermediate particle size diameter of D_{84} values was found to be 0.80 mm and 0.65 mm for LHS and UHS, respectively. Roughness coefficients (n) were also similar for each site and ranged between 0.08 – 0.1.

Table 4. Highest water levels observed for two stream sites

Year	Stream site	
	LHS	UHS
	cm	cm
Year 1	101	-
Year 2	139	72
Year 3	117	55

The LHS stream site showed more consistency in the geometric relationships assess and likely due to the fact that more field discharge values were recorded for LHS compared to UHS site (Table 4). These relationships were obtained as an initial evaluation of the five methods of discharge estimation. Products and the sum of m and p exponents should close to 1 to be consistent with Q=A.V equation (Lawrence, 2007), and I found that LHS provided more reasonable estimates for those exponents which were close to 1. In contrast, the sum and products of m and p deviated from 1 in some cases in the UHS stream site (Table 5) and likely because of the fewer field measurements. This suggests a minimum (at least 10) number of field discharge measurements required to determine the consistency of these geometric relationships.



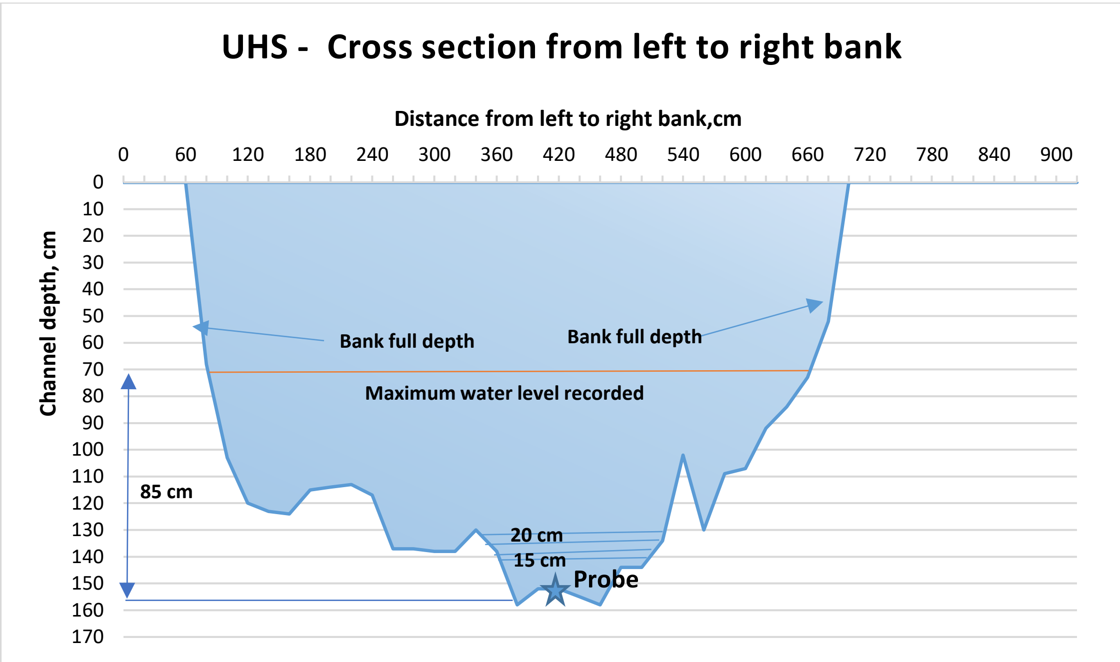


Figure 4. The shape of the channel cross section measured from left to right bank, looking upstream and the location of the water level probe for LHS (top panel) and UHS (bottom panel) plotted using the field measurements.

Table 5. Geometry factors and their relationships used in assessing continuity via the Q=AV equation for UHS and LHS stream sites.

Site **Method**	**Exponents** m	**Exponents** p	**Coefficients** k	**Coefficients** o	**Product of coefficients** k*o	**Sum of exponents** m + p
UHS						
Curve fitting	0.145	0.855	0.850	1.177	1.000	1.000
FC method	0.217	0.468	0.364	0.433	0.158	0.685
Jarrett's method	0.855	8.421	8.421	1.177	9.907	9.276
Combined method	0.305	0.695	0.780	1.283	1.000	1.000
Ferguson method	2.446	1.446	2.887	2.869	8.282	3.892
LHS						
Curve fitting	0.446	0.554	2.537	0.394	1.000	1.000
FC method	0.624	0.330	0.448	2.092	0.938	0.954
Jarrett's method	0.306	0.694	0.699	1.430	1.000	1.000
Combined method	0.617	0.383	0.505	1.980	1.000	1.000
Ferguson method	0.440	0.560	0.341	2.937	1.000	1.000

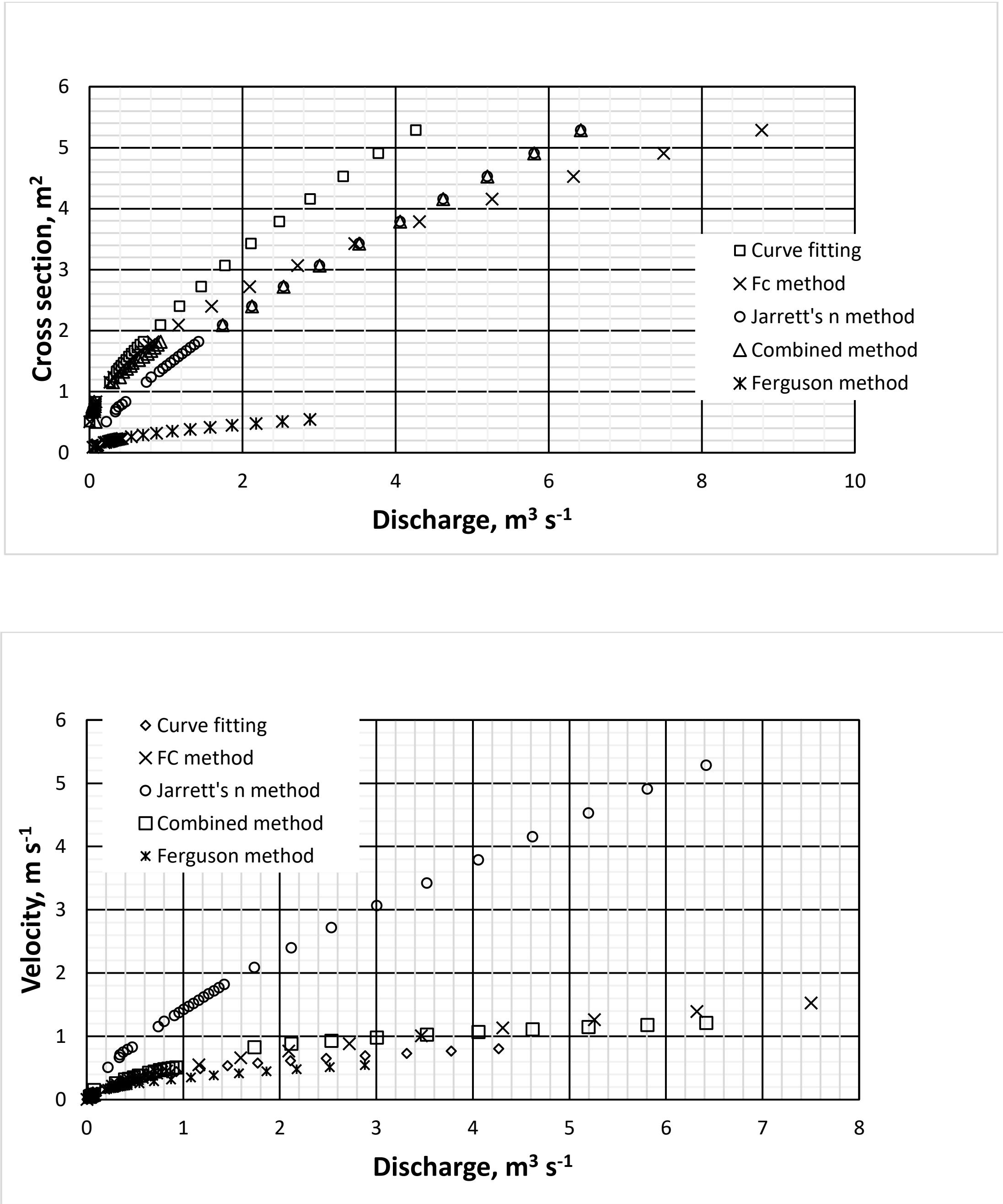


Figure 5. Relationships for cross-sectional area vs. discharge (upper panel) and velocity vs. discharge (lower panel) derived from all five methods used in the study and applied to the LHS site.

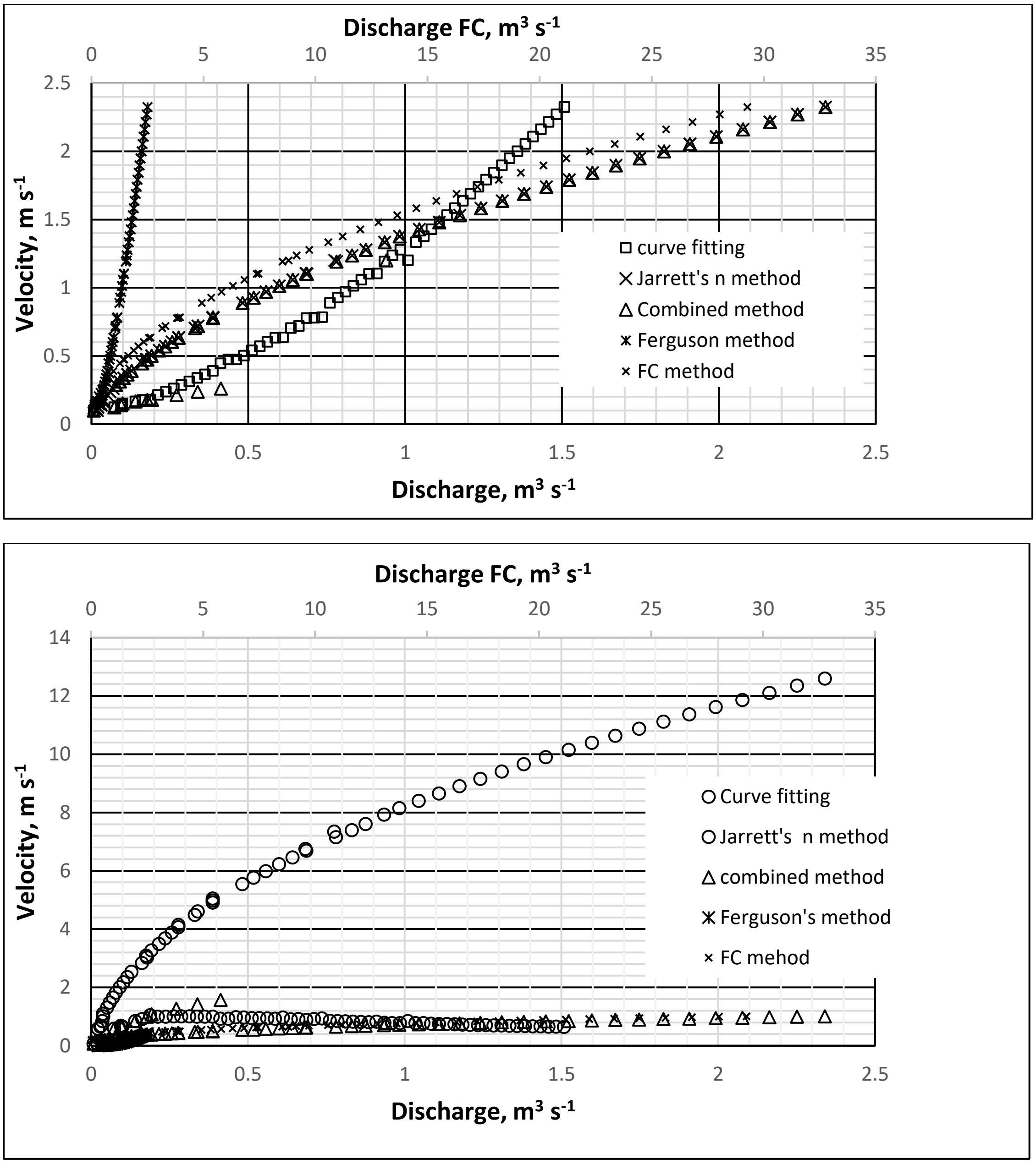


Figure 6. Relationships for cross-sectional area vs. discharge (upper panel) and velocity vs. discharge (lower panel) derived from all five methods used in the study and applied to the UHS site.

2.3.2. A comparison of the five approaches for generating stream rating curves

There were several practical issues identified when a curve fitting method was applied to the datasets collected for the two streams sites studied. Results revealed that the curve fitting method relied more on the geometric features of these stream sites when developing an adequate rating curve covering the full range of gauge heights.

Both curve fitting and the FC methods provided reasonable estimates of discharge at lower gauge heights and performed well for gauge heights up to 50 cm with an error of ± 6% (Table 6). However, discharge estimates using the FC method yielded significantly higher values for 1 m gauge height which are above those made in field measurements (Table 7). In fact, the FC method provided the highest discharge estimates among all the methods as exhibited in data for a gauge height of 1 m in both LHS and UHS stream sites (8.78 $m^3 s^{-1}$ and 29.2 $m^3 s^{-1}$, respectively; Table 6). In contrast, the lowest estimates of discharge among all the methods were obtained using Ferguson's method which estimated discharge to be 2.88 $m^3 s^{-1}$ and 0.6 $m^3 s^{-1}$ at LHS and UHS, respectively, at the same gauge height (Table 6). The predicted total discharge value of Jarrett's method was 1.37 $m^3 s^{-1}$ for 50 cm gauge height whereas actual field discharge measurement was 0.83 $m^3 s^{-1}$. Moreover, Jarrett's method provided reasonable estimates for 1 m gauge height for both LHS and UHS stream sites (6.45 $m^3 s^{-1}$, 3.36 $m^3 s^{-1}$, respectively) and the predicted total discharge of 6.45 $m^3 s^{-1}$ for that gauge height was 26%

lower compared to the discharge value of FC method for the same gauge height (Table 5).

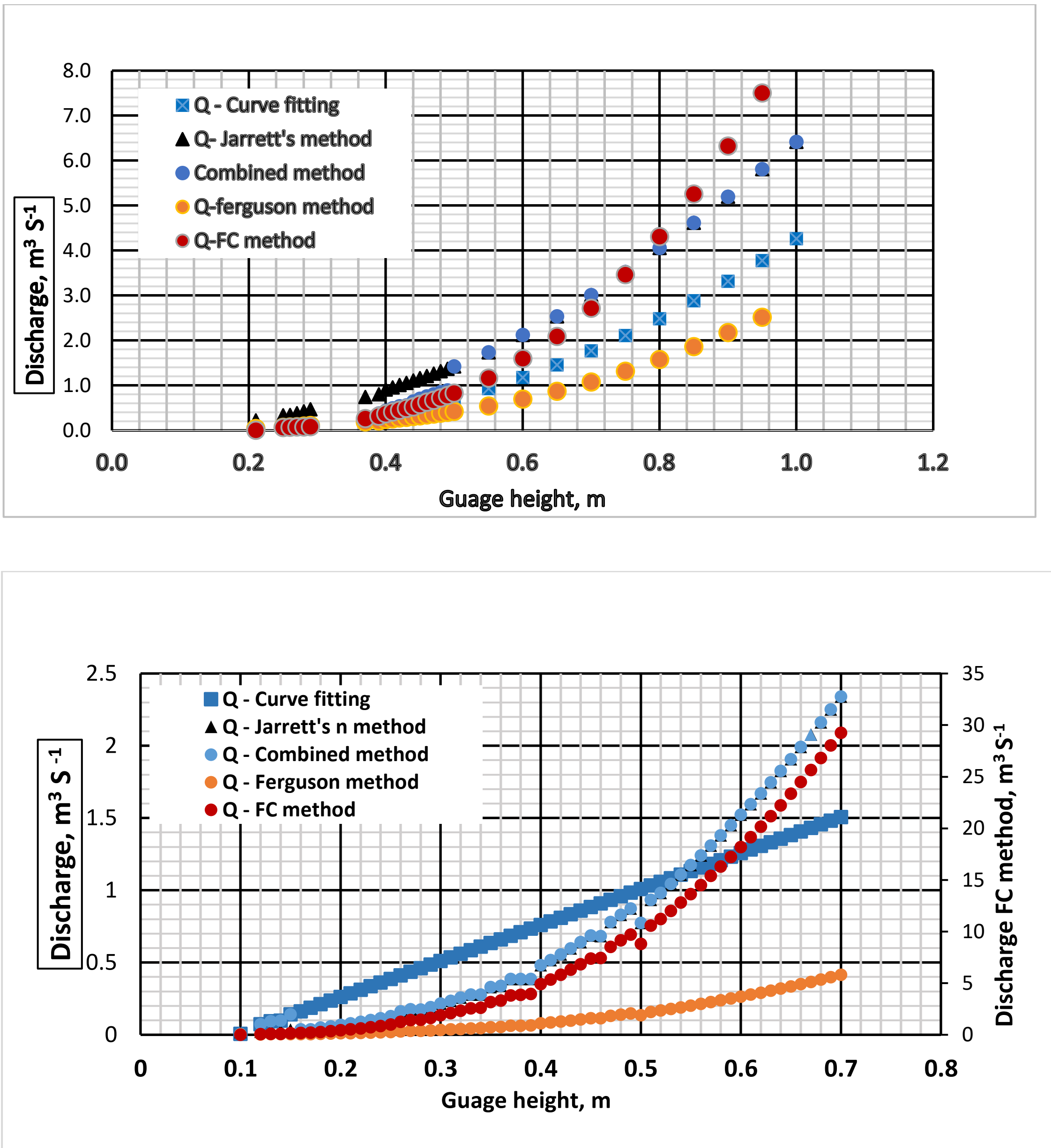


Figure 7. Discharge estimates made using each of the five methods assessed in this study for the LHS (top) and UHS (bottom) stream sites. Discharge estimates using the flow conveyance (FC) method for the UHS stream site was plotted in the secondary y-axis in the bottom panel).

Table 6. The predicted total discharge for a gauge height of 50 cm, which corresponds to a field measurement of 0.83 $m^3 s^{-1}$, as determined by each of the five different approaches and provided with percent error.

Method	Error for 50 cm of field discharge	
	Field discharge for 0.83 $m^3 s^{-1}$	
	$m^3 s^{-1}$	Percent error
Curve fitting	0.71	-12%
FC method	0.78	-5%
Jarrett's method	1.37	+65%
Combined method	0.75	-9%
Ferguson method	0.55	-33%

Table 7. Estimated discharge for a 1 m gauge height at both the LHS and UHS site using the five different prediction methods used. No direct measurements of discharge are available for this gauge height.

Method	Estimated discharge for 1 m of the upper section of the rating curve	
	LHS	UHS
	$m^3 s^{-1}$	$m^3 s^{-1}$
Curve fitting	4.26	2.25
Flow conveyance (FC method, Manning's eq)	8.78	29.26
Jarrett's n method	6.45	3.36
Combined method (FC and Jarrett's)	6.45	3.36
Ferguson method (Ferguson, 2006)	2.88	0.60

Considering the results obtained for FC and Jarrett's methods, both (Table 6 and 7) performed well for specific sections of the rating curve, but poorly for other sections of the rating curve. As an example, the FC method better estimated discharge for the lower part of the rating curve while Jarrett's method performed well for the upper section. With these observations, I was encouraged to define lower and upper portions of the rating curve separately as each method performed well for different sections (Table 6 and 7).

Lack of field information that was constrained in the FC method which may lead to poor performance in predicting discharge at higher gauges. Only the channel cross section and wetted perimeter were used to constrain discharge, which appears to be not enough to describe the features of rocky headwater streams. The main limitations of the FC method to lower gauge heights are the changes in channel roughness coefficient values, and slope gradients with rapidly changing depths specially in headwater streams at high energy levels (Jarrett, 1987; Roberts et al., 2016). However, by constraining more field data such as roughness coefficient values may have added more accuracy to Jarrett's estimates as shown in these results. As recommended by Jarrett's (1987), constraining roughness coefficient values in predicting discharge for high energy streams was more successful and appropriate in comparison with the FC method.

A combined rating curve using each of these two approaches but applied to specially defined parts of the rating curve and resulted in a better overall rating curve then either was able to provide individually. A discharge of 0.75 $m^3 s^{-1}$ was predicted by the combined rating curve for 50 cm gauge height which was roughly a similar estimate to the measured discharge of 0.83 $m^3 s^{-1}$ within the 9% error associated with this field measurement for that gauge height. However, the discharge for the upper limit of the rating curve for 1m gauge height was reported as 6.45 $m^3 s^{-1}$ by combined rating curve. As suggested by Jarrett's (1987), I used Eq. 7 to calculate appropriate Manning's roughness coefficient values for higher gauge heights. By addressing the uncertainty associated with stream roughness values, especially for rocky headwater streams at high flows conditions, could reduce the error associated with high discharge predictions.

However, several challenges should be addressed when considering a combination of methods such as determining the lower and upper limits of a rating curve. As an illustration, in this study, I used FC estimates up to field discharge measurements and Jarrett's method was used for the upper section of the rating curve. Ferguson's method provided the overall lowest discharge predictions for the entire catchment. Those discharge values for total catchment area were 2.66 $m^3 s^{-1}$ and 0.6 $m^3 s^{-1}$ for LHS and UHS, respectively. Methods that constrain more features that describe the channel geometry seemed to be more accurate in establishing adequate rating curves and Jarrett's and Ferguson's methods appeared to be in successful in predicting those values by consuming more field data in these two approaches. Selecting the best approach for predicting discharge and accessing the predicted discharge values, have become quite challenging. However, comparing the predicted runoff ratios with regional runoff ratios seems more reasonable in accessing the estimated values.

Section 2.3.3 Evaluation of the combined rating curve and the rating curve based upon Ferguson's method for estimating discharge based upon runoff ratios.

Table 8. Discharge, runoff and runoff ratio estimates for Combined and Ferguson methods

Year	Total discharge×10^6(m^3)		Total runoff×10^2 (mm)		Runoff ratio	
	Combined	Ferguson	Combined	Ferguson	Combined	Ferguson
Year 1	33.75	15.00	25.22	9.74	1.70	0.83
Year 2	27.39	11.57	20.47	8.70	1.54	0.79
Year 3	19.02	9.27	14.21	6.93	1.03	0.55

The rating curve based upon Ferguson's method appeared to have provided the most reasonable estimates of discharge for the two stream sites investigated here as compared with the combined rating curve. The combined rating curve generated relatively high total discharge estimates of 3.4 X 10^7 m^3, 2.8 X 10^7 m^3, and 1.9 X 10^7 m^3 in years 1, 2 and 3, respectively, as compared with the rating curve based upon Ferguson's method (Table 8). These higher discharge values resulted in unreasonable runoff ratios of 1.70, 1.54 and 1.03 for year 1, 2 and 3, respectively, as compared with the regional runoff ratio of 0.65 as derived from the historical precipitation data for the area obtained from Environmental Canada (Humber river, station number 02YL012, Environmental Canada). Total annual discharge for the total catchment area, as measured at LHS estimated using Ferguson's method were lower than those were estimated using the combined rating curve method and found to be 1.3 X 10^7 m^3, 1.2 X 10^7 m^3, 9.3 X 10^6 m^3 for year 1, 2 and 3, respectively. Consequently, the runoff ratios based upon discharge determined via Ferguson's method rating curve were lower and more comparable to regional values (0.83, 0.77 and 0.55 for year 1, year 2 and year 3, respectively; Table 8). When comparing the runoff ratios, the combined rating curve appeared to over predict the discharge. However, these results reveal that the Ferguson method provides more realistic discharge and runoff estimates.

Streams like headwater streams show huge variations in their geometry and morphology and these variations were addressed in Ferguson’s method by considering grain size data, slope, hydraulic radius, and flow resistance. During high flows, flow resistance does much impact when submerged areas provide extra resistance to flow and by applying flow resistant data including grain size in Ferguson’s method, provided meaningful discharge estimates. Total resistant to flow was described as relative roughness in Ferguson approach. Therefore, grain size data acted as distinguishing feature Ferguson’s method adding more information of flow resistance in this method.

2.3.4. Error analysis for predicted Ferguson’s rating curve

Discharge estimates were based on Ferguson estimates using Ferguson equation depending on D_{84} value, R, and slope measurements (R equals to cross section divided by wetted perimeter and I considered the cross section as it was measured at the field sites). Each of the factors that were used to determine any discharge measurement using Ferguson’s equation has some level of uncertainty and requires a reasonable estimate of uncertainty. The best estimate of the uncertainty of each of those factors was to determine the full range of potential values estimated for discharge. However, I was unable to determine the full uncertainty associated with the rating curves developed for each site. Therefore, I determined the variation observed in each of the factors used and calculated the full range of known variation for the discharge estimates made.

I determined the percentage variation of each of the three parameters required in the Ferguson equation by considering the number of measurements, measured at the field at different times. For example, the slope of the LHS stream site was measured two times during 2017 and 2018. I used the most recent values for my rating curve developments and the percentage variation of the most recent values to the previous measurement was calculated to determine the % variation of the slope. Following the same procedure, the percentage variation in the measured D_{84}, slope, and cross section measurements of the channel was calculated as 10%, 13%, and 10%, respectively. For each parameter, I calculated the highest and the lowest possible values using the above percentages. I considered each possible combination (eight combinations in all) of the highest and lowest values within the stage-discharge equation to calculate eight sets of Ferguson discharge values. Then, obtained the highest and lowest possible discharge values to calculate the full range of discharge estimates (Fig. 4). After determining the highest and the lowest values, I came up with three stage-discharge relationships, one for the original relationship and the other two for the highest and the lowest estimates. Range of variation for predicted Ferguson's discharge estimates were increased from lower gauge heights to higher gauge heights. Upper and lower bounds of that variation for LHS and UHS sites were presented in table 9.

Table 9. Range of variation values for Ferguson discharge estimates

Gauge height	**LHS**			**UHS**		
	Original	Lower	Upper	Original	Lower	Upper
	m^3s^{-1}	m^3s^{-1}	m^3s^{-1}	m^3s^{-1}	m^3s^{-1}	m^3s^{-1}
50 cm	0.55	0.43	0.59	0.29	0.26	0.36
100 cm	2.88	2.28	3.13	0.92	0.84	1.12

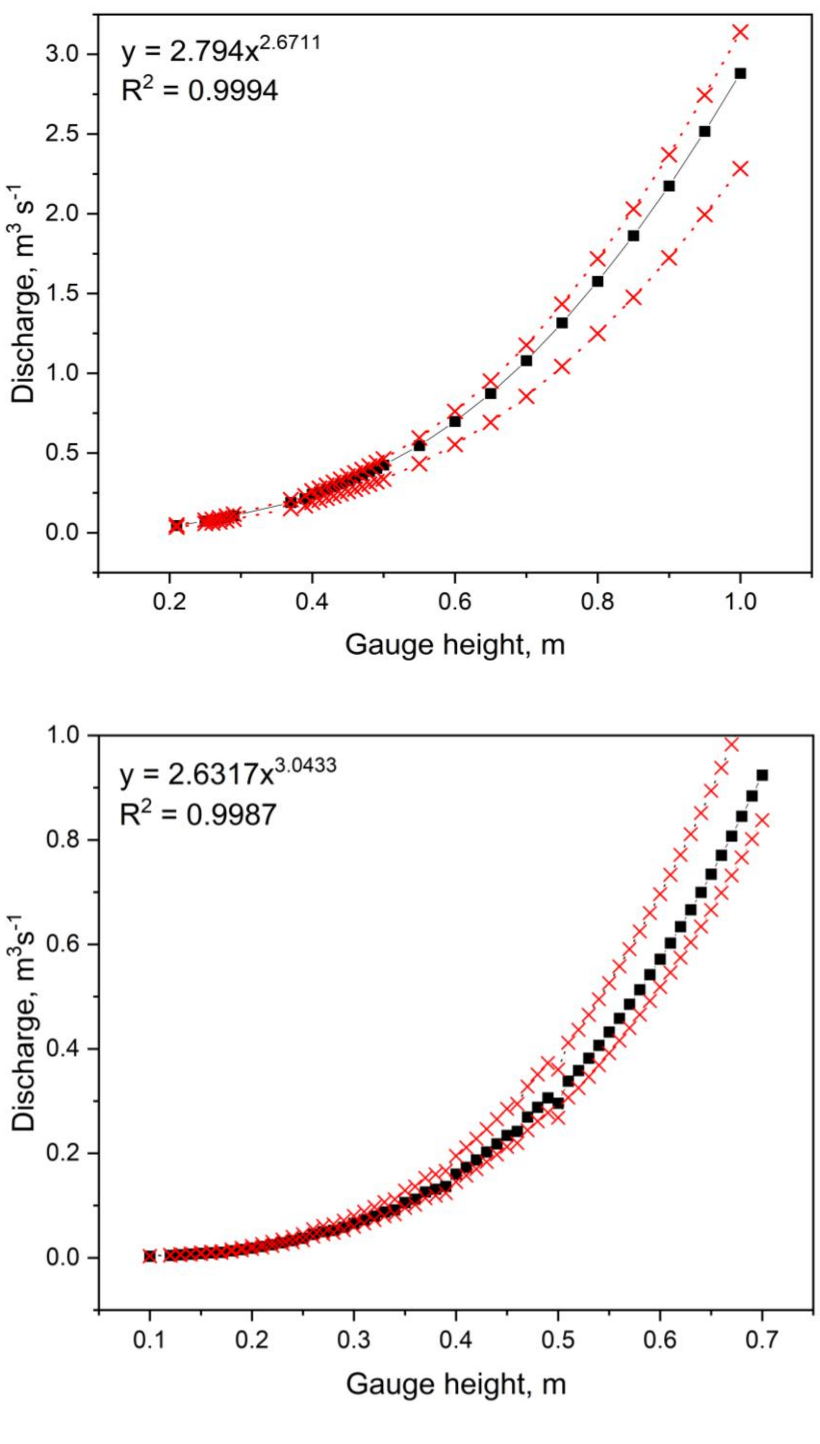


Figure 7. Ferguson rating curve for LHS (top) and UHS (bottom) stream sites. Lower and upper uncertainty ranges were plotted in red color.

2.4. Implications of Ferguson's resistance equation applied to the rocky mountainous stream

Flow velocities of these small streams depend on various factors which describe the geometry of the stream channel. Therefore, methods that constrain more field information have become more successful to predict indirect discharge values, especially for first-order rocky streams. Ferguson's method was developed based on field parameters related to the geometry of the channel. However, more data were required and constrained for this approach such as the gradient of the channel, grain size data, hydraulic radius, and average depth values which can be used to improve rating curve development (Table 3).

Headwater streams are heterogenous in geometry and morphology and complex systems to predict indirect discharge values (Mcdonnell *et al.*, 2007). As suggested by many authors, flow resistant act as a prominent factor in controlling streamflow (Prestegaard, 1983; Ferguson, 2007; Roberts *et al.,* 2016). Ferguson method was far more accurate and reliable in means of extending rating curves as Ferguson's method fulfilled the requirement of more field parameters like flow resistant to predict discharge in heterogenous headwater streams. Results of this study revealed that having information to better exploit the geometry for stream sites such as rocky headwater stream help to develop rating curves to estimate discharge. However, there were difficulties in capturing the heterogeneity to obtain a more accurate estimate of discharge for different stream networks such as small and higher order scale streams (Mcdonnell *et al.*, 2007). Heterogeneity influence greatly on the water flux variations on those scales of streams where there is a large variation of the channel geometry and roughness. However, in order to develop methods of estimating indirect discharge in headwater streams with steep gradients, hydraulic jumps, plunge pool sequences etc., new methods should be developed to assess the channel geometry and the roughness more accurately.

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Chapter 3.0 Annual and seasonal variations in water fluxes of a boreal forested headwater stream.

Abstract

Terrestrial to aquatic fluxes are reliant on hydrology which in turn is controlled by climate, morphology and land cover features of the boreal landscape. Climate change impacts the hydrology of boreal landscapes through changes in snow accumulation and melting which impact stream discharge and solute transport to aquatic environments. In this chapter, I investigated seasonal and interannual variations of stream water and DOC fluxes for an experimental watershed in western Newfoundland, Canada. Two stream sites were gauged to understand impacts of the shifts in precipitation from snow towards rain with the earlier melt of snowpack on water and DOC fluxes and to quantify the contributions of two different watershed components to water and DOC export. Three contrasting winters, one characterized by a more continuous snowpack (year 1), the other two by a mild and a major mid-winter melt events (years 2 and 3, respectively), exhibited annual water exports of 1.5×10^7 m^3, 1.2×10^7 m^3 and 9.5×10^6 m^3, respectively, suggesting reduced export associated with mid-winter melting of the snowpack. Consistent with this trend the annual runoff ratios for years 1, 2, and 3 were 0.71, 0.67, 0.46, respectively. The hillslope region, representing roughly half of the catchment area contributed over 80% of the total catchment water and DOC, whereas the upper low gradient pond and wetland dominated portion exhibited much lower yields of water and DOC. The short

snowmelt period contributed most of the annual water (50%) and annual DOC (>65%) export from the whole catchment. Differences in annual and season discharge and runoff ratios in two landscape components indicated that the hillslope portion is more responsive to the reduction in the snow fraction and increased mid-winter melt events whereas these changes had no impact on discharge or runoff ratios for the upper low gradient portion of the catchment. This indicates that reductions in the spring snowmelt events with climate warming have the potential to increase the proportion of discharge and solute sources from low gradient portions of these boreal landscapes. Therefore, results from this work suggest that reduction in hillslope lateral flow water inputs associated with a reduced spring snowmelt event may be associated with (1) reductions in overall stream water discharge, and (2) increases in low gradient pond or wetland contributions to the aquatic environment in these boreal landscapes.

3. 1 Introduction

The timing, distribution and form of precipitation are susceptible to climate change and can greatly influence the timing and magnitude of streamflow, runoff and the delivery of solutes from terrestrial to aquatic environments. Such changes can have important implications for aquatic ecosystems and biogeochemical feedbacks. For example, current estimates of terrestrial to aquatic fluxes of carbon indicate this flux (>2 Pg C globally) equals or exceeds the net exchange of CO_2 between atmosphere and land or sea (Drake et al., 2016; Cole et al., 2007; Tranvik et al., 2009; Raymond et al., 2016). These global estimates are derived indirectly from the sum of evasion and burial of carbon in freshwaters and large river transport of carbon to the sea making it difficult to directly assess how these fluxes are controlled. More direct measurements of terrestrial to aquatic fluxes are possible through measurements within small headwater stream where the close connection between the landscape and aquatic environment exist. Therefore, the knowledge of headwater stream water fluxes can greatly enhance our understanding of controls of carbon flux as well as other impacts on the aquatic environment such as nutrient delivery.

The boreal landscape contains globally significant stores of terrestrial carbon (Raymond and Saiers, 2010; Vucetich et al., 2000) and contain extensive hydrologic networks that control the terrestrial to aquatic exchange of carbon and other elements. These high latitude landscapes are also quite vulnerable to climate change and have, therefore, been the subject of the study of terrestrial and aquatic stores and fluxes of carbon and nutrients (Vucetich *et al.*, 2000; Algesten *et al.*, 2003; Tranvik *et al.*, 2009; Norris *et al.*, 2011; Laruelle *et al.*, 2014; Lepistö et al., 2014; Tranvik et al., 2014). Some such studies have revealed increasing trends in dissolved or total organic carbon (DOC or TOC, respectively) concentration and export in some higher latitude streams and rivers in regions of North America and Europe over the past few decades (Driscoll *et al.*, 2003; Skjelkvåle *et al.*, 2005; Evans *et al.*, 2006; Roulet, 2006). Drivers of these trends remain elusive, however, in most cases there are links to environmental and climate change impacts on watershed components and hydrology. Some studies have linked trends in DOC export to trends in air temperature (Freeman et al. 2001); occurrence of severe droughts (Worrall et al., 2004), recovery from acid deposition (Freeman et al., 2004), and land management trends (Worrall et al. 2003). Some of the mechanisms suggested by these trends include changes in availability of organic matter sources, such as increased productivity or soil decomposition enhanced by warming (Freeman et al., 2004; Fenner et al., 2007), while many others are likely consequences of changing discharge amounts, timing and flow paths (Tranvik and Jansson, 2002; Hejzlar et al. 2003). For example, in regions underlain by permafrost, warming trends coincide with reduced DOC export attributed to increased infiltration of DOC rich surface waters into mineral rich soil layers where adsorption immobilizes DOC (Striegl et al., 2005). In contrast, increasing DOC concentrations or export could be associated with increased discharge in watersheds of mixed wetland and forest composition as a consequence of (1) increased connectivity between saturated areas or wetlands and catchment areas contributing to flow (Boyer et al. 1997; Stieglitz et al. 2003; Creed et al. 2003), and

(2) preferential flow through organic rich surface soils with macropore flow bypassing mineral soil adsorption processes (Hornberger et al. 1994; McGlynn and McDonnell, 2003). Therefore, changes in magnitude, form and timing of precipitation associated with climate change can create shifts in streamflow magnitude and timing which can impact terrestrial to aquatic solute transport.

Decreasing snow as a proportion of total precipitation identified over the past few decades as a result of climate change (Buttle et al*.,* 2005; Devito *et al.,* 2005), likely impacts catchment hydrology and hydrologic processes relevant to boreal stream transport (Glenn and Woo, 1997; Taylor and Pierson, 1985). Reduction in snow deposition and, therefore, melt may in fact reduce late spring and summer stream flows in catchments with snow dominated hydrology, hence the snow resources are melting early as a consequence of climate change (Groisman *et al.*, 2004; Barnett *et al.,* 2005). With earlier melt of snow reducing the magnitude and changing the timing of spring melt discharge (Barnett *et al.,* 2005; Schelker *et al.*, 2013; Irannezhad *et al.*, 2017). The increased winter runoff reduced stored snow, earlier and reduced spring snowmelt runoff, coincide with reduced summer streamflow volumes observed in mountainous western North American streams (Mote, 2003; Hamlet et al., 2005). In fact, a continental scale analysis indicates that reductions in the proportion of precipitation as snow with climate change are associated with reduced streamflow (Berghuijs et al. 2014).

The response of boreal streams to changing precipitation regime will depend upon key elements that make up the boreal landscape (i.e. wetlands and upland forests; Köhler et al., 2008). Boreal landscapes show a wide range of landscape heterogeneity leading to increased complexity in watershed hydrology (Mcdonnell *et al.*, 2007). Headwater streams in recently deglaciated landscapes usually drain from ponds or lakes, and wetland before traversing through gradient hillslopes (Devito *et al.,* 2005; Schelker *et al.*, 2013). Infiltration processes and runoff pathways are different in low relief catchments with wetland and pond regions compared to regions with steep hillslopes connected directly to stream channels (Dunne, 1978; McDonnell et al., 2007). Lakes and ponds can facilitate evaporation loss of water to the atmosphere but largely act as storage increasing the residence times within the catchment (Tranvik et al., 2009) particularly in more humid boreal regions.

Catchment topography, soil characteristics, and vegetation influence hydrological flow paths (Sivapalan *et al.,* 1987; Mcdonnell *et al.*, 2007). Landscape features like forests can influence the magnitude, timing, and response of streams to precipitation. For example, forests can impose a seasonal influence on stream hydrology through evapotranspiration during summer and reduction in snowpack, via canopy interception, over winter which can reduce spring melt runoff. Furthermore, forest harvesting may change both evapotranspiration demand runoff generation processes (Devito *et al.,* 2005; Creed *et al.*, 2013) and those landscape practices may potentially increase groundwater tables after harvesting (Schelker et al., 2012) which could lead to increases in stream baseflow discharge (Sørensen *et al.*, 2009; Hornbeck *et al.,* 2011).

Different landscape positions react to hydrology differently as hillslopes are more hydrologically connected to streams whereas the low relief wetlands can be more hydrologically isolated. Hillslopes often contain well drained soils and during high wetness periods, hillslope soils can be highly transmissive and can provide significant quantities of water to near stream sites (McGlynn and McDonnell, 2003; Peters et al., 1995). This is probably why in wet years the TOC export from forested catchment areas responds more readily to precipitation events in snow-free periods whereas wetlands or mire catchments exhibit no real seasonal response (Kohler et al. 2008). Areas with higher gradients promote lateral flow depending upon soil type where low permeable soils create low infiltration and can promote lateral flow through upper more permeable layers (Jencso et al., 2009). Macropore flow within the subsurface of forested soils plays an important role in developing preferential flow paths in the delivery of water and solutes (Tsuboyama et al., 1994; Tsukamoto and Ohta, 1988). Upland hillslopes of boreal landscapes contain materials ranging from moss and organic soil layers to rocky mineral soils and fractured bedrock. In general, hydraulic conductivity decreases with depth; these hydraulic conductivity contrasts may be more extreme in regions with thick moss and organic horizons. Contrasts in hydraulic conductivity between soil layers on combined with steep gradients likely promote shallow lateral flow which will influence both water fluxes and solute concentrations and composition (Mcglynn *et al.*, 2004)**.**

Though contrasts in responses of low relief regions, including wetlands in stream riparian regions, and upland forest regions to snowmelt or rain events are likely important to solute transport in boreal regions they remain poorly understood. Steep hillslopes, particularly those with soil macropores, can convey water downslope through the shallow subsurface (Tsuboyama *et al.*, 1994; Mcdonnell *et al.*, 2007), however, the degree to which they connect and contribute to streamflow depends greatly upon the water balance and storage capacity aspects regulated by region climate, geology and topography. Likewise, boreal ponds and wetlands located in upper, poorly developed watersheds can be hydrologically disconnected from stream channels, except during extreme events (Jencso *et al.*, 2009; Creed et al., 2018). Wetlands and ponds act as water storage components which makes these regions response differently to rain events or early snowmelt as a result of changing climate (Devito *et al.,* 2005). What is unclear is to what extent saturated overland flow dominants hydrologic pathways in these low relief boreal features relative to steeper forested hillslopes particularly during snowmelt periods.

The contrasting role that boreal wetland and forest regions can have on catchment biogeochemical fluxes, and in response to hydrologic events, underscores the need to better understand intra and inter-annual variation in boreal stream hydrology associated with these elements. For example, alterations of timing and the form of the winter precipitation patterns linked to climate change, for example, appear to be a potentially significant feature controlling increased river DOC fluxes in a high latitude watershed (Huntington et al., 2016)**.** A better understanding of the hydrologic responses of boreal headwater streams and their solute fluxes to variation in precipitation regime could aid in our understanding of the potential drivers behind such trends.

This study focusses on a small forested watershed in a mesic forested region of the NW Atlantic where snow is a significant portion of the annual precipitation. Previous research in this watershed indicates that water can move rapidly through surface organic layers and carry significant DOC fluxes within forested landscapes (Bowering et al., 2019). Pynn's Brook Experimental Watershed is located in the mesic NW Atlantic boreal region of Canada where both soil DOC fluxes and runoff are relatively high. The watershed is characterized by low gradient areas near streams in the upper watershed and steep hillslopes near streams in the lower watershed that are dominated by black spruce forests (Moroni et al., 2010). These differences in watershed morphology provide a wide range of environmental conditions to investigate the spatial and temporal variations of water and solute fluxes in boreal headwater streams.

In this study, I examine water flux variations at two locations in a small boreal forest stream over 3 years with contrasting in winter precipitation regimes. Stream gauges are located to separate the upper low-gradient pond and wetland-dominated upper catchment area from the lower catchment that contains mainly steep hillslope regions near an incised stream with no noticeable floodplain. I assessed how seasons and precipitation regimes impacted water flux contributions from these morphologically distinct components of the catchment. The main objectives of this chapter were to; (1) Quantify the annual, interannual and inter site water flux variability among the three years of different winter precipitation; (2) Assess the differences in water fluxes and runoff ratios associated with the low gradient upper catchment and the and steeper lower catchment regions; and (3) Use relationships between stream discharge and DOC concentrations to estimate the stream dissolved organic fluxes (DOC) to assess the potential influence of water the flux variations studied on biogeochemical processes occurring in the experimental watershed.

3.2. Methods

3.2.1 Study area

The study was conducted in the Horseshoe Brook located within the Pynn's Brook experimental watershed (PBEWA) on the west coast of the island of Newfoundland, in eastern Canada (having coordinates of 48° 53' 14" N, 63° 24' 24" W). The Horseshoe Brook is a headwater stream with a total catchment area of approximately 11.37 km^2. The climate in this area is characterized by winters with consistent snow coverage for approximately four to five months of the year beginning from December to April. The annual average precipitation is 1095 mm that shows year to year variation, and the average annual daily mean temperature is 4 ^{0}C (Environment Canada Climate Normals, Deer Lake Airport 1981-2010). The Horseshoe Brook is a second order stream draining an upper catchment area (3.84 km^2) of small headwater lakes and wetlands and a similarly sized (7.52 km^2) with steep hillslopes (Fig.9). Non-wetland regions are largely forested by black spruce (*Picea maiana*) and balsam fir (*Abies balsamea*) forests varying from mature (>80 years of age) to harvested within the last 10-15 years.

The upper catchment is characterized by a relatively low relief area having four ponds of varying sizes. The lower region hereafter referred to as Hillslope which is characterized by steep hillslopes which were investigated separately from the upper catchment area by subtracting the UHS fluxes and runoff from those determined for LHS.

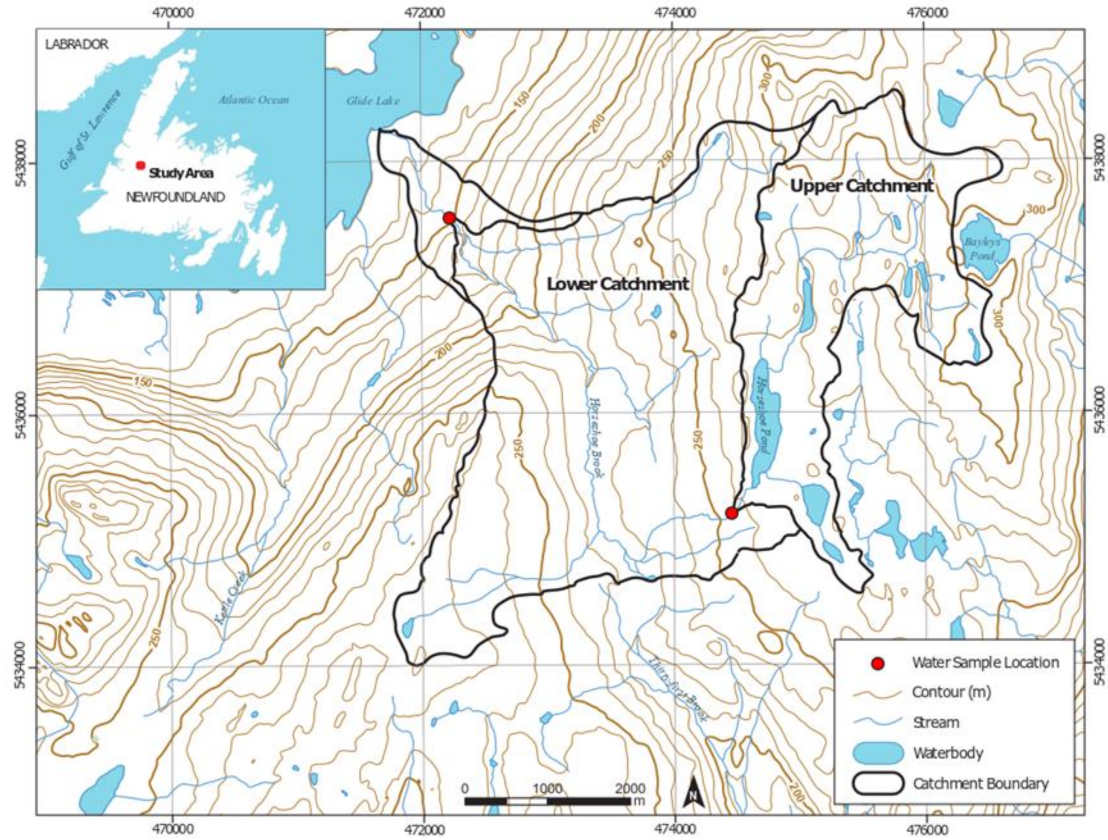

Figure 9. A map of the catchment area for the Horseshoe Brook located near Deer Lake, western Newfoundland and Labrador, Canada. The two study stream sites, Lower Horseshoe (LHS) and Upper Horseshoe (UHS), are indicated by red circles and approximate catchment area delineated for each site shown in black.

3.2.3. Stream site data collection

Dataset

Continuous stream water level, conductivity and temperature data were collected at the LHS over three separate years between October 2014 to October 2018 using in situ pressure transducer (Onset U20 water level logger; MA, USA) and conductivity dataloggers (Onset U24 conductivity logger; MA, USA). The same equipment was installed, and data were collected at the UHS except these did not commence until October 2016. Continuous water level was recorded at 30 min time interval (continuous water levels were captured for 15 min time intervals for May 2017 to Oct 2017 time period) and conductivity data was captured for every 30 min time intervals except for October 2014 to October 2015 (conductivity data were recorded for every 15 min intervals). Downloaded data from the probes were atmospheric pressure corrected using the atmospheric pressure data downloaded from atmospheric pressure loggers located at each stream site using another pressure transducer (Onset RX3000 data logger, MA, USA). However, there were no atmospheric probes were installed at the sites in 2014, therefore, the source for atmospheric correction for 2014 was the downloaded atmospheric pressure data from Environmental Canada at the Deer Lake station (station number - 02YL007) (49°13'00" N, 57°24'00" W) located approximately 20 km away. The existing barometric data, collected at the LHS (from 2016 October to 2018 October), were found to be well correlated with the Deer Lake Station barometric pressure data (r^2= 0.882, p<0.0001).

Water samples were collected for DOC at each field site from 29th March 2017 to 8th of May 2018 on a seasonal basis. In total, 15 and 7 samples for DOC measurements were collected from the LHS and UHS stream sites, respectively. Samples were filtered using 0.45µm PES filters and the filtrate (15 mL) was stored in 24 mL percombusted glass vials with clean Teflon lined caps. These samples were acidified to a pH of 2 using 150 µL of 20% v/v HPLC grade H_3PO_4 and stored, dark and at 4˚C until the analysis was done. Analysis for DOC was conducted using Shimadzu TOC-V analyzer. To obtain a rough estimate of stream DOC concentration over the course of the study period, a multiple regression relationship was established between the measured DOC concentrations and the continuous datasets of water level, temperature, and conductivity monitored in each site. The variance of the estimates was evaluated from the R-square value obtained for the model.

3.2.4. Methods used to estimate water fluxes from continuous gauge height data

Several methods were discussed in chapter one to extend the field discharge measurements beyond the base flow levels and generate a useful rating curve for these rocky headwater stream sites. Velocity for high flows was estimated using bed roughness (grain size) and water surface gradient to obtain resistance coefficients (Ferguson, 2016). These velocity estimates were combined with the channel area measurements to obtain discharge-gauge height relationships (known as a rating curve for both stream gauge locations. These rating curve relationships were used with the continuous water level data downloaded from the probes installed at the field sites to estimate water discharge for the period of this study.

The following equations were used to evaluate flow resistance and velocity in order to determine the rating curves to extend the field measurements. Hydraulic radius and grain size data or intermediate diameter of streambed particle distribution that equals or exceeds that of 84 percent of the particles (D84 values) were used to compute the velocities and the discharge values. The flow resistance was related to relative roughness (d/D_{84}) and calculated by dividing the average depth (d) by D_{84} for boulder bed streams as described in Ferguson (2007). Shear stress was calculated using the following equation:

$$u^* = g^{0.5} \cdot R^{0 \cdot 5} \cdot S^{0 \cdot 5} \qquad \text{Eq. (10)}$$

Where u^* is shear stress, g is gravitational acceleration, R is hydraulic radius (cross section/wetted perimeter), and S is the stream gradient. Hydraulic radius was determined by cross sectional area divided by wetted perimeter for each gauge height and shear stress was calculated by multiplying the square root of g, R, and S (Eq. 10). Velocities were then determined using the following equation from the calculated u^* values from Eq. (10).

$$\frac{u}{u^*} = 2 \cdot \left(\frac{d}{D_{84}}\right) \qquad \text{Eq. (11)}$$

The resistance parameter, u/u^* is estimated from relative roughness using Ferguson's equation (Eq. 10), which uses the depth, d to particle size, D_{84} ratio. D_{84} is the particle with an intermediate diameter that equals or exceeds that of 84 percent of the particles. Values of u/u* for each water level was determined using Eq. (10) and each u/u* value was multiplied by their associated u* value to determine u. The value of *u* was then multiplied by the channel cross section to estimate discharge estimates. Using the discharge calculations described above I generated a representative rating curve for the full range of gauge heights. Then a stage-discharge relationship was generated to calculate the continuous discharge estimates using the continuous water level data.

Interannual, intra annual, and inter-site variations water fluxes were calculated from the estimated continuous discharge values. Although I am unable to determine the full uncertainty associated with the rating curves developed for each stream site, I determined the variation observed in each of the factors used and calculated the full range of known variation for the discharge estimates made. Upper and lower bounds of discharge values were determined based on the calculated range in rating curves for each site using that full range of variation in the measured field values of D_{84} value, R, and slope (R equal to cross section divided by wetted perimeter and I considered the cross section as it was measured at the field sites). Full ranged variation in each of the factors that were used in Ferguson discharge estimations was used to obtain the full range of discharge values as predicted from this approach. These upper and lower values are all included in the results in order to capture what variation I was able to assess in these datasets.

I determined the percentage variation of each of the three parameters required in the Ferguson equation by considering the number of measurements, measured at the field at different times. For example, the slope of the LHS stream site was measured two times during 2017 and 2018. I used the most recent values for my rating curve developments and the percentage variation of the most recent values to the previous measurement was calculated to determine the % variation of the slope. Following the same procedure, the percentage variation in the measured D_{84}, slope, and cross section measurements of the channel was calculated as 10%, 13%, and 10%, respectively. For each parameter, I calculated the highest and the lowest possible values using the above percentages. I considered each possible combination (8 combinations in all) of the highest and lowest values within the stage-discharge equation to calculate 8 sets of Ferguson discharge values. Then, obtained the highest and lowest possible discharge values to calculate the full range of discharge estimates (Fig. 10). After determining the highest and the lowest values, I came up with three stage-discharge relationships, one for the original relationship and the other two for the highest and the lowest estimates. Finally, these relationships were used to calculate the highest and the reduced discharge estimated on an annual and seasonal basis to determine the upper and lower bounds of these estimates. The upper and lower estimate of DOC were calculated by considering both the full range in rating curves, and thereby discharge, and the multiple regression model predictability for DOC concentration.

Runoff ratios were determined using the total precipitation for the catchment area. Tipping bucket rain gauge (RST Instruments Model TR-525) was installed and used to monitor local rainfall at the sites. Data from the tipping bucket was compared with the regional rainfall data reported by Environmental and Climate Change Canada (Station 02YL007, Environmental Canada) at Deer Lake A station (49°13'00" N, 57°24'00" W), located approximately 20 km from the study catchment, and found to be well correlated (R^2=0.882, $p<0.0001$). Regional precipitation data obtained from the Deer Lake A station was used for the runoff calculations. Estimated discharge values were converted to mm of runoff by dividing the catchment area in order to directly compare those estimates with the precipitation. Obtained discharge values were then divided by the precipitation values to calculate the annual and seasonal runoff ratios. The total annual runoff ratio values were compared with the regional runoff ratios for the Humber River, a 5th order river with a catchment area of 7,860 km^2 (Humber Village Station, 02YL001, Environment and Climate Change Canada 2010-2018). The study catchment is located within the Humber River watershed and 45 km from the Humber Village discharge station. Runoff ratios were also obtained for a nearby 3rd order stream, South Brook, as another reference to determine the validity of the calculated runoff values from each approach assessed here. The South Brook discharge station (02YL004, Environment Canada) is located approximately 30 km from the study catchment and captures the runoff from a steep catchment with a total area of 58.5 km^2.

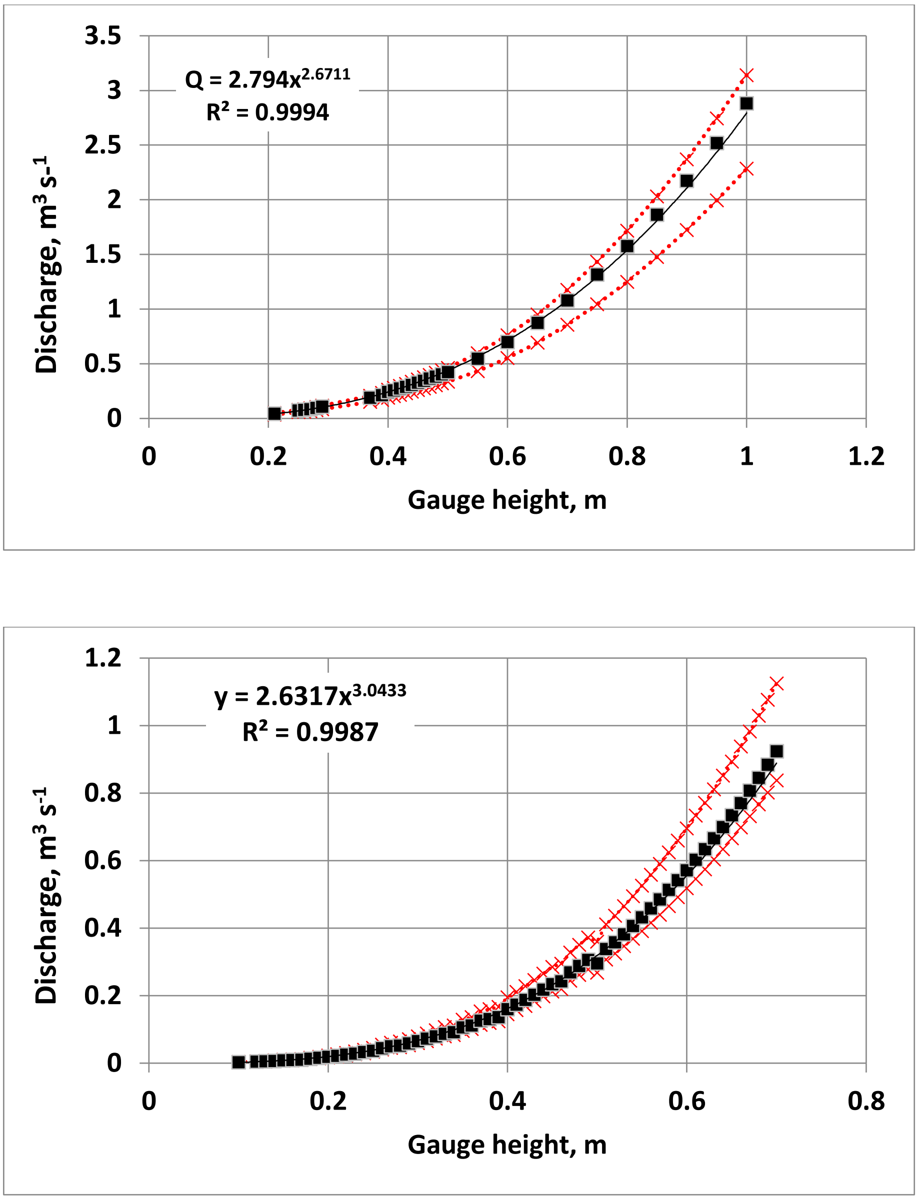

Figure 10. Rating curve extended using Ferguson's resistance equation as described in Chapter 2 and applied here to obtain stage-discharge relationship to calculate discharge estimates for Lower Horseshoe Site (LHS; top) and Upper Horseshoe Site (UHS; bottom) stream sites. The upper and lower estimates are indicated in red color (x).

3.2.5. Estimating DOC fluxes

Infrequent grab samples were not sufficient to develop relationships between dissolved organic carbon (DOC) concentration and discharge. Such relationship(s) are required to estimate stream DOC fluxes, Therefore, relationships between DOC concentration and conductivity were developed and used along with continuous conductivity measurements to predict DOC concentration was generated from the multiple regression analysis of with continuous water level and the conductivity data as independent variables and DOC concentration as a dependent variable. Two multiple regression equations were developed, one for UHS and one for LHS with the available DOC concentrations. The form of the equations is:

$$Y_1 \text{ (DOC concentration)} = a+ \beta_1 * \text{(water daily volume)} + \beta_2 * \text{(daily conductivity)} \quad \text{Eq. (12)}$$

Table 10. Regression analysis results obtained for UHS and LHS sites

Stream site	β_1	β_2	C	R - square
UHS	5.3×10^{-4}	1.3×10^{-1}	8.8×10^{-1}	0.6
LHS	3.5×10^{-6}	-5.0×10^{-1}	7.0×10^{0}	0.5

Acknowledging the poor predictive power indicated by the low R^2 values for these relationships (Table 10), the application of these relationships was used only to obtain a broad estimate of DOC flux, without the precision needed to decipher seasonal or even annual differences. Furthermore, the upper and lower estimates of the DOC fluxes were determined using these variations in the DOC concentration estimates combined with the upper and lower estimates of the water fluxes.

The annual hydrograph was divided into four seasonal designations that are based on hydrological observation including the generally consistent timing of evapotranspiration shutdown, snowpack development and spring snowmelt among the three years studied. The same periods were selected for each season within each study year to avoid significant changes for the beginning and ending and length of each season for each year. The start of the fall in any given year was based upon the evapotranspiration shut down defined as the date when the largest increase in groundwater water level occurred in the absence of significant rainfall (Table 11). The end of fall and the start of winter was designated as the average date of the occurrence of the first snowfall that marked the beginning of the formation of a continuous snowpack. Within each hydrological year the fall period was split into an early fall and late fall period because each study year started in fall in part to avoid splitting up the snowpack period into separate years. However, both fall periods were summed up to calculate the water flux and runoff for each fall period. The spring melt period represents the average period from when the snowpack begins to decrease to when no snow was completely lost. The summer period consists of some of the calendar year spring period and ends when evapotranspiration ends (Table 11).

A 30-day gap in the continuous water level dataset for LHS occurred during the spring melt to early summer period (May 8 to June 9, 2015). Therefore, a double mass curve was developed and used to estimate discharge for this period. Discharge data downloaded from the South Brook station, located approximately 30 km away from the catchment was used with existing LHS data from this study in order to build a double mass relationship (Station number - 02YL004, Environment Canada). Year 2 was selected to obtain the relationship where there was a complete continuous data record for the entire year, and because 2018 data are not yet available for South Brook. Cumulative discharge from LHS was plotted against the cumulative discharge of South Brook discharge (Fig. 11). The resultant relationship (R^2=0.99; Fig. 11) was then applied to year 1 to fill the missing data gap in order to have a complete spring melt period for year 1. I noted that the period that I extrapolated was the shaded area in Fig. 11.

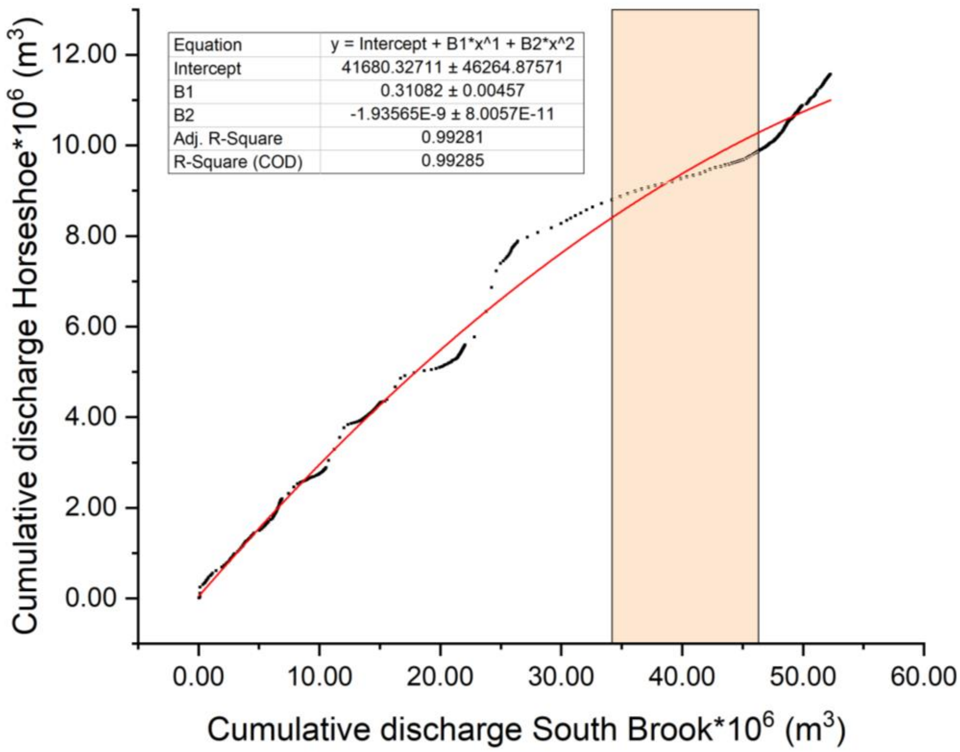

Figure 11. The double mass curve using 2017 data of cumulative discharge for the South Brook station (Station number -02YL004, Environment and Climate Change Canada) and the Upper Horseshoe (UHS) Site from this study. The shaded area depicts the dates of the year when this relationship (shown in red) was applied to predict LHS discharge for the missing spring melt data in 2014.

Table 11. Annual and seasonal precipitation in mm for three different years of this study (Year 1 - 2014 October to 2015 October, Year 2 – 2016 October to 2017 October, and Year 3 – 2017 October to 2018 October).

Season	Time period	**Year 1** (Continuous snowpack)	**Year 2** (Small mid-winter melts)	**Year 3** Large mid-winter melts
Late fall	Oct 6 - Dec 10	311	318	192
Winter	Dec 11 - April 6	544	463.9	562
Spring Melt	April 7 – May 29	119	145.8	185
Summer	May 30 – September 9	321	238	279
Early fall	September 9- October 5	286	120	286
Total		1581	1286	1505
% snow		42	34	34
Number of snow melting events		Spring melt only	4 mid-winter + spring melt	One major mid-winter+ spring melt

3.3. Results

The total annual precipitation in each of the three study years (1581 mm, 1286 mm, and 1505 mm for year 1, year 2, and year 3, respectively) was higher than the mean annual precipitation for the period 1981-2010 (1095 mm; source – Average normal, Deer lake A station, Environmental Canada). The proportion of the snow to the total precipitation was greatest in year 1 and represented 42%, 34%, and 34% of total annual precipitation for year 1, year 2, and year 3, respectively (Table 11). The three winter periods represented the highest proportion of the total annual precipitation (an average of ~35%) while the lowest proportion was recorded during the short spring melt periods. However, there was an increased proportion of precipitation as snow in year 3 during the spring melt period indicating a shift in snowfall compared to the other two years (Fig. 12).

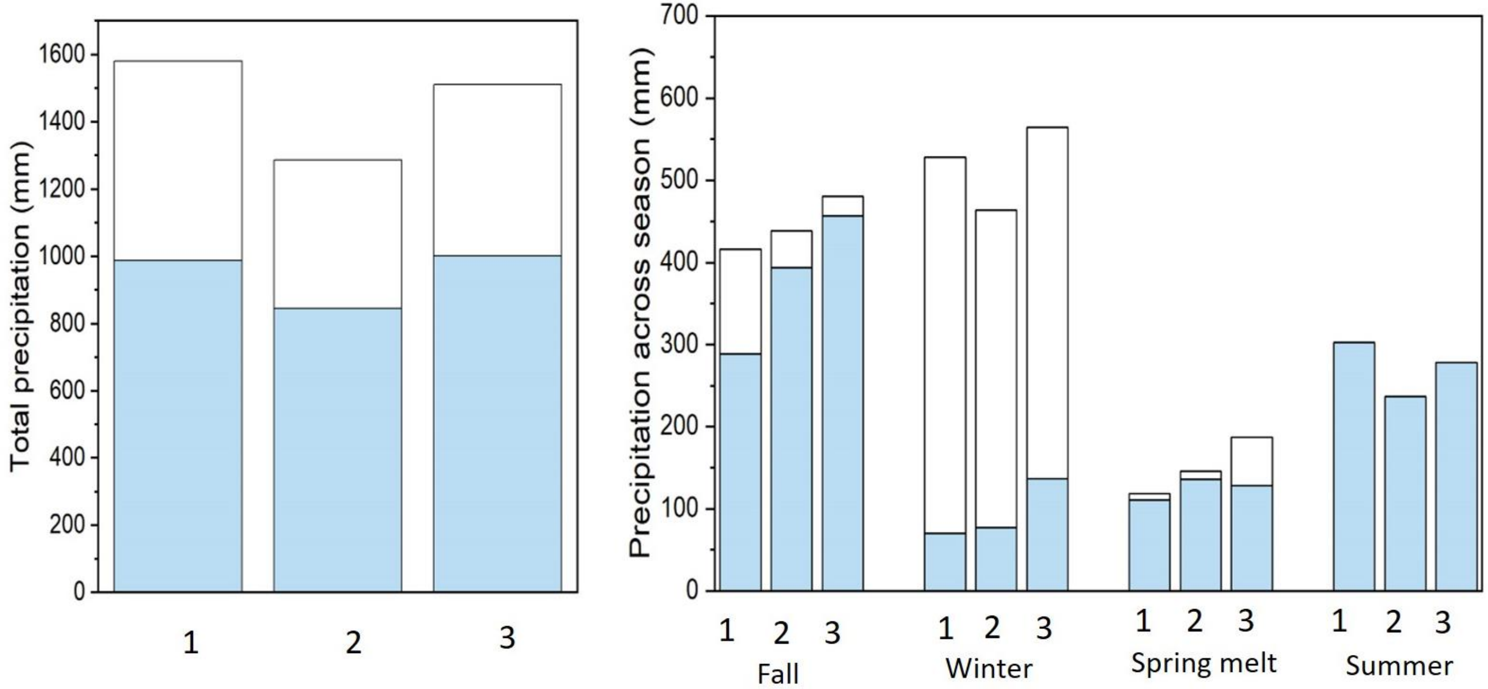

Figure 12. Annual precipitation in mm for each of the three study years (1, 2, and 3 values represent year 1 (October 2014 – October 2015), year 2 (October 2016-October 2017), and year 3(October 2017-October 2018)). Total snow was presented in white color and blue indicates the total rain (Source – Deer lake A (station number - 02YL007) Environmental Canada website). Total snow was converted to mm of rainfall using a 10:1 maritime snow to water equivalent conversion factor (Sturm et al. 2010).

Table 12. Annual and seasonal discharge (×10^6 m^3) for the Lower Horseshoe Site (LHS). The lower and upper bounds of the discharge estimates are presented within brackets. Each season was defined considering the sequence of the seasons occurred within each hydrological year.

Season	Year 1 (m^3)	Year 2 (m^3)	Year 3 (m^3)
Late fall	3.3 (2.6-3.6)	2.1 (1.7-2.3)	1.3 (1.1-1.5)
Winter	3.7 (3.0-4.1)	3.5 (2.8-3.8)	3.7 (3.7-4.1)
Spring melt	3.8 (3.3-3.9)	4.1 (3.2-4.4)	1.9 (1.5-2.1)
Summer	2.4 (1.9-2.6)	1.2 (1.0-1.3)	1.1 (0.9-1.2)
Early fall	1.8 (1.4-2.0)	0.7 (0.6-0.8)	1.4 (1.1-1.5)
Total	15.0 (12.3-16.1)	11.6 (11.6-12.6)	9.5 (7.5-10.3)

Table 13. Annual and seasonal discharge (×10^6 m^3) for the Upper Horseshoe Site (UHS) and the hillslope portion of the Horseshoe Brook catchment determine from the LHS discharge minus the UHS discharge. The lower and upper bounds of discharge estimates are presented within brackets next to each estimate.

Season	Year 2		Year 3	
	(m^3) UHS	(m^3) Hillslope	(m^3) UHS	(m^3) Hillslope
Late fall	0.28 (0.26-0.34)	1.81 (1.40-1.93)	0.10 (0.09-0.12)	1.28 (0.98-1.34)
Winter	0.13 (0.12-0.16)	3.36 (3.05-3.65)	0.23 (0.20-0.27)	3.51 (2.75-3.79)
Spring melt	0.25 (0.23-0.31)	3.82 (3.00-4.13)	0.26 (0.24-0.32)	1.61(1.29-1.79)
Summer	0.06 (0.06-0.08)	1.14 (0.90-1.24)	0.12 (0.11-0.15)	0.99 (0.77-1.06)
Early fall	0.06 (0.06-0.08)	0.65 (0.51-0.70)	0.16 (0.14-0.19)	1.21 (0.94-1.30)
Total	0.79 (0.72-0.97)	10.78 (8.46-11.65)	0.86 (0.78-1.05)	8.61 (6.73-9.28)

Table 14. Annual and seasonal runoff ratios for the Lower Horseshoe Site (LHS). The lower and upper bounds of the runoff ratio estimates are presented within brackets.

Season	Year 1	Year 2	Year 3
Late fall	0.78 (0.62-0.85)	0.49 (0.39-0.53)	0.52 (0.41-0.56)
Winter	0.51 (0.40-0.55)	0.56 (0.45-0.61)	0.47 (0.39-0.47)
Spring melt	2.35 (2.09-2.47)	2.09 (1.66-2.27)	0.76 (0.62-0.84)
Summer	0.55 (0.45 -0.60)	0.38 (0.30-0.41)	0.30 (0.28-0.32)
Early fall	0.47 (0.37-0.51)	0.44 (0.35-0.48)	0.36 (0.28-0.38)
Total	0.83 (0.69- 0.97)	0.79 (0.65-0.85)	0.55 (0.45-0.59)

Table 15. Seasonal runoff ratio for the Upper Horseshoe Site (UHS) and the hillslope portion of the Horseshoe Brook catchment. The lower and upper bounds of runoff ratios estimates are presented within brackets next to each estimate.

Season	Year 2		Year 3	
	UHS	Hillslope	UHS	Hillslope
Late fall	0.13 (0.12-0.16)	0.86 (0.66-0.92)	0.08 (0.07-0.09)	0.97(0.76-1.05)
Winter	0.04 (0.04-0.05)	1.09 (0.86-1.18)	0.06 (0.05-0.07)	0.94 (0.74-1.02)
Spring melt	0.26 (0.24-0.32)	3.95 (3.10-4.27)	0.22 (0.19-0.26)	1.35(1.05-1.45)
Summer	0.04 (0.04-0.05)	0.72 (0.57-0.78)	0.07 (0.06-0.08)	0.53 (0.42-0.57)
Early fall	0.08 (0.07-0.10)	0.81 (0.63-0.87)	0.08 (0.07-0.10)	0.64 (0.50-0.68)
Total	0.09 (0.08-0.11)	1.26 (0.99-1.36)	0.09 (0.08-0.11)	0.86 (0.67-0.93)

Snow as a proportion of the total precipitation was reduced from 42% in Year 1 to 34% in years 2 and 3 (Fig. 12). The three years of gauge height and discharge data between 2014 – 2018 represented three different hydrological years based on the variation of the precipitation (Table 11). Year-1 was a hydrological year with a consistent and continuous snowpack and large spring snowmelt without any real melting events during the winter. Year-2 and year-3 were distinct from year-1 as they included a couple of storms or melting events during the winter period. In Year-2, 4 small melting events occurred throughout the winter. In Year-3 a nearly total snowpack removal occurred as a result of an increase in temperature and a large rain event that occurred in mid-winter (January 10th - 30th).

Highest total annual discharge of 1.5×10^7 m^3 ($1.2–1.6 \times 10^7$ m^3) was observed in year 1 corresponding to the year with the highest annual precipitation. Total annual discharge for year 2 and year 3 were 1.2×10^7 m^3 ($9.2 \times 10^6 – 1.3 \times 10^7$ m^3) and 9.5×10^6 m^3 (7.5×10^6 -1.0×10^7 m^3), respectively (Table 3.3; Fig. 14 A and C), despite the similarly in total precipitation between year 1 and year 3 and lower precipitation in year 2. Runoff and the runoff ratios reflected a similar decreasing trend from year 1 to year 3 where the total runoff ranged from 1118 mm to 708 mm from year 1 to year 3. Annual runoff ratios for year 1, year 2, and year 3 were 0.83, 0.79, 0.55, respectively (Table 14). Discharge hydrographs exhibited differences in the magnitude and number of discharge events associated with snowmelt events among the three study years consistent with the precipitation patterns. No significant mid-winter melt discharge events and only one late fall discharge event were observed in year 1. In year 2 a small late fall, larger early and two mid-winter discharge events occurred, and in year 3 two large mid-winter discharge events occurred as a result of the complete snowpack loss (Fig. 13). Winter and spring melt periods contributed over 50% of the total discharge for each year. The largest proportion of total discharge observed in fall occurred in year 1 (35%) while the proportion of total discharge observed in fall for years 2 and 3 was 24% and 28%, respectively. An increasing trend in the proportion of the total discharge occurring in winter was observed from year 1 to year 3 consistent with increases in the magnitude of winter snowmelt events. The highest proportion (40%) of total discharge observed in winter was recorded in year 3 when the largest mid-winter melt occurred, and the lowest value was reported for year 1 (25%) when snowmelt was confined to small events in early winter.

Despite the short spring snowmelt period, the discharge was generally comparable among seasons within each study year with the exception of summer which was typically lowest (Fig. 14 B and C). Fall and summer discharge was greatest in year 1 while winter discharge was similar among years despite the large variation in snowmelt event number and magnitude. It was the spring melt discharge that appeared most impacted by the variation in mid-winter melt events with year 3, when the most significant mid-winter melt occurred, having exhibited the lowest spring melt discharge.

There were no significant differences in annual runoff ratios for year 1 and year 2 while year 3 had the lowest annual runoff ratio (Fig. 14 C). The highest seasonal runoff ratio was generated during the spring melt in year 2 (280 mm, 304 mm for year 1, and year 2, respectively), but was then found to be lowest in year 3 when it was only 144 mm. Expectantly, spring melt runoff ratios for the three years were the greatest among all the seasons (Table 16; Fig. 14 D). However, there was a decrease in spring melt and summer runoff ratios from year 1 to year 3. For example, the spring melt runoff ratio, ranged from 2.35 to 0.76 for year 1 through 3 while the summer runoff ratio ranged from 0.55 to 0.30.

3.3.1 Variation in discharge and runoff ratio between the upper, low gradient catchment and the lower, hillslope-dominated catchment.

The UHS gauge drains the upper catchment, an area of low gradient, whereas the LHS gauge includes the upper catchment and adds the lower basin steep hillslope segments to the catchment. The lower hillslope component is quantified by evaluating the discharge differences between the two stream gauges. These topographically different portions of the catchment were associated with different annual and seasonal and runoff volumes and ratios.

Despite the similar catchment areas, the hillslope dominated part of the catchment contributed over 10-fold more discharge as compared to the upper lower relief area of the catchment. Considering the total catchment area, the hillslope region contributed more than 80% of the total catchment discharge (Table 13; Fig. 15). Annual water volumes of 1.1×10^{7} m^3 and 8.6×10^{6} m^3 were observed for the hillslope region in year 2 and year 3, respectively, while discharge values of 8.0×10^{5} m^3 and 9.0×10^{5} m^3 were observed for the UHS region for year 2 and year 3, respectively (Table 13; Fig. 15). Greatest annual runoff ratios were found for the hillslope region and those values decreased from year 2 to year 3 which is consistent with the precipitation in each given year (Table 15; Fig. 15).

The largest difference in water volumes and runoff ratio between the two catchment components were observed in winter (Table 15; Fig. 15). The upper low relief area exhibited increased discharge and runoff ratio in winter and summer with the largest mid-winter snowmelt occurring in association with higher annual precipitation in year 3. While these patterns were not observed in the hillslope region where the large mid-winter melt lead to no change in winter discharge and runoff ratio but a much-reduced spring melt discharge and runoff. Furthermore, the years with greater precipitation coincided only with an increase in fall runoff ratio for the hillslope portion of the catchment.

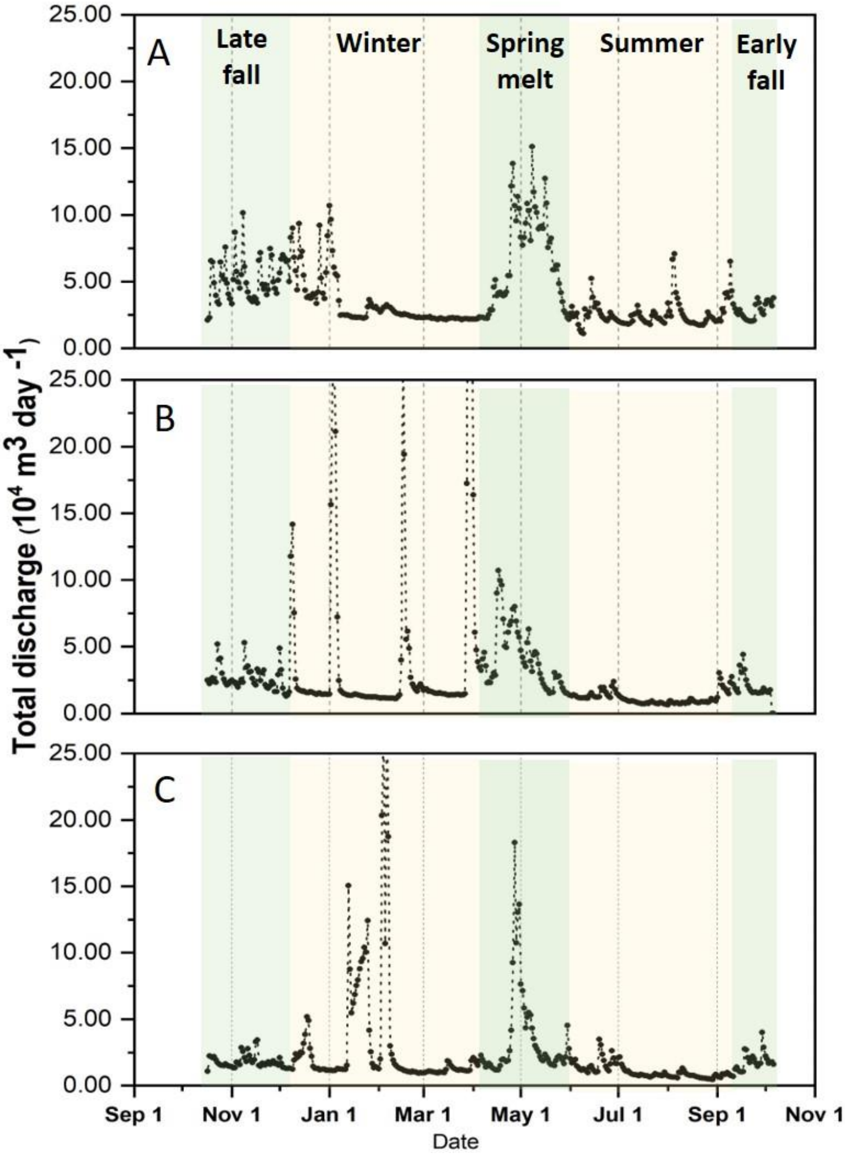

Figure 13. Continuous water flux results for all the three study years (A – year 1, B- year 2, C- year 3) obtained for the lower horseshoe (LHS).

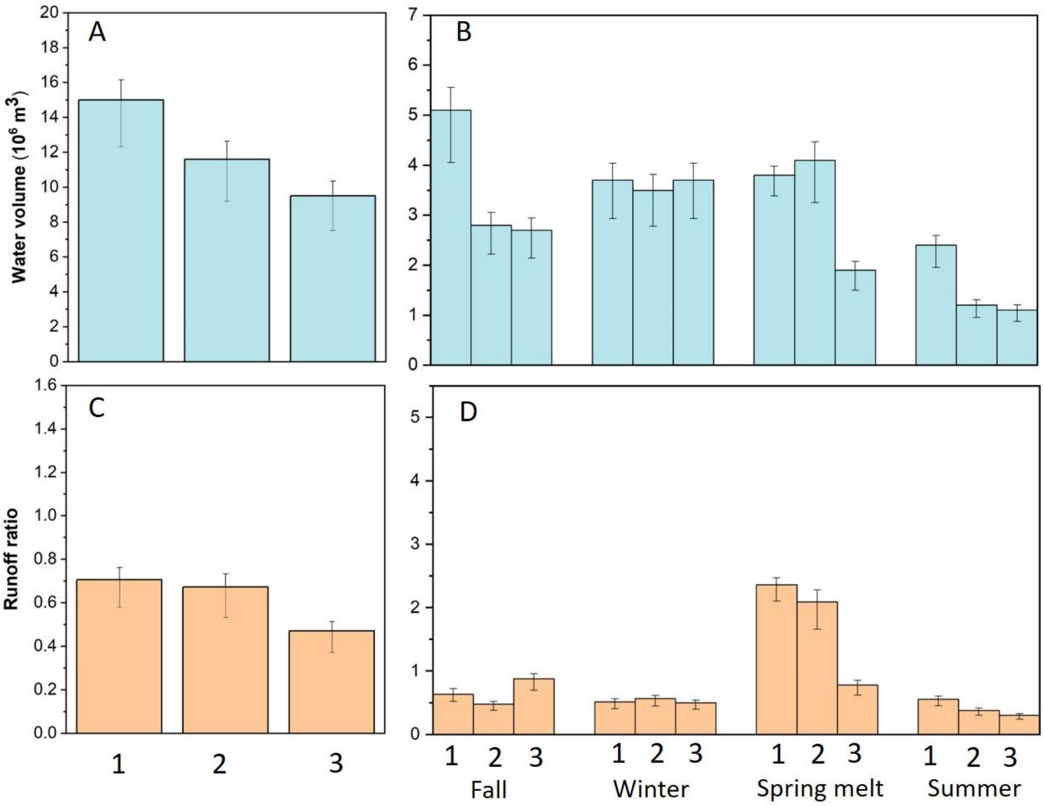

Figure 14. Total annual (A, B) and seasonal (B, D) lower horseshoe (LHS) water volume and runoff ratios in three years. Error bars represent the lowest and highest estimates from the actual value. (1, 2, and 3 values represent year 1 (October 2014 – October 2015, year 2 (October 2016 – October 2017), and year 3 (October 2017 – October 2018).

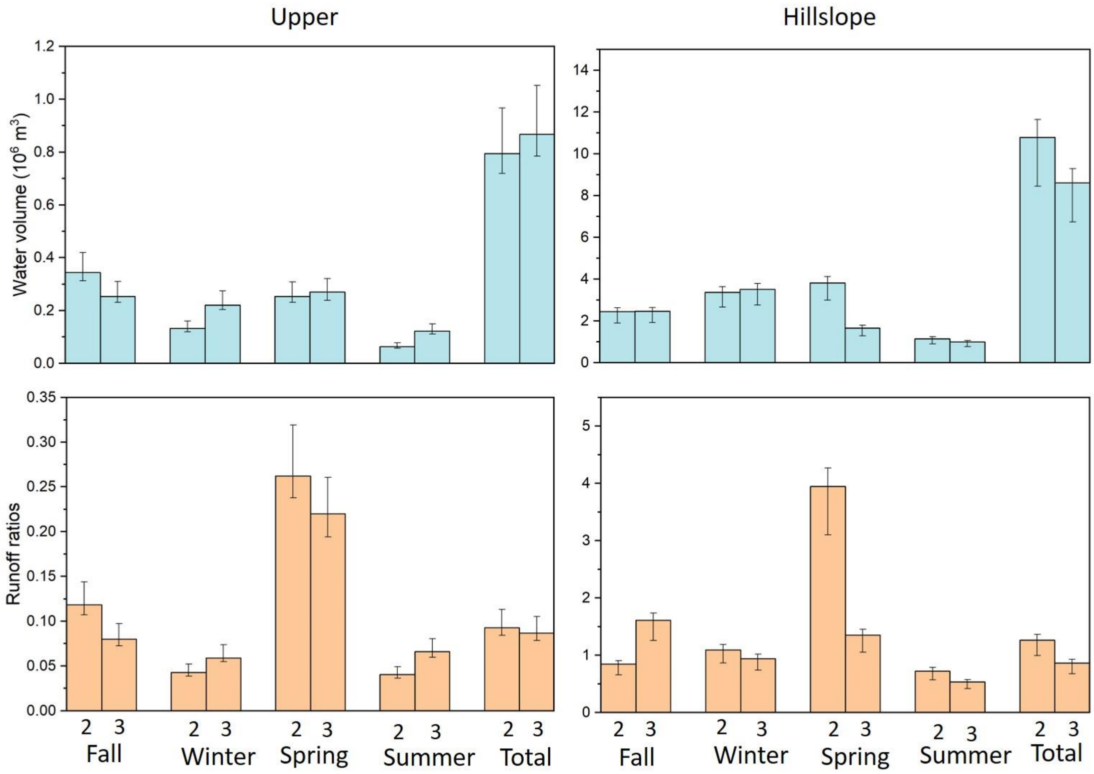

Figure 15. The seasonal water volume and runoff ratios for the upper low relief (left panels) and the lower hillslope dominated (right panels) catchment component. Error bars represent lowest and highest estimates of each value (1, 2, and 3 values represent year 1 (October 2014 – October 2015, year 2 (October 2016 – October 2017), and year 3 (October 2017 – October 2018).

3.3.2. Estimated DOC fluxes among years and for the two contrasting catchment areas

There was no significant difference ($P < 0.05$) in the estimated total DOC flux among three years (Fig. 16) and the range in the estimates based upon what uncertainty we could constrain was typically ~50% of the mean value. The magnitude of the range in the DOC flux estimates likely result from the very low resolution in sampling effort for DOC concentration and precludes any possibility of detecting differences among years, seasons and perhaps even components of this study catchment. DOC fluxes for the catchment were estimated to be 5.18 g C m^{-2} (2.87- 8.70 g C m^{-2}), 5.29 g C m^{-2} (2.48 -8.13 g C m^{-2}), and 4.58 g C m^{-2} (2.16-7.02 g C m^{-2}), for year 1, year 2, and year 3, respectively.

The largest differences in seasonal DOC fluxes among the three years occurred in the late fall period: values ranged from at 1.38 g C m^{-2} (0.65 - 2.11 g C m^{-2}) in year 1 to 0.56 g C m^{-2} (0.27-0.86 g C m^{-2}) in year 3. The winter and spring melt DOC fluxes exhibited the highest seasonal fluxes, despite the relatively short spring melt flux period, while the summer seasonal DOC flux was lowest throughout the study. Fall DOC fluxes were, however, quite comparable with both winter in spring. Greatest relative variation in seasonal DOC flux occurred in summer with Y3 exhibiting the lowest. There was no detectable variation in DOC fluxes for winter among the three years while the largest spring melt DOC flux was reported for year 2 while the lowest was recorded in year 3, the year with the largest mid-winter snow melt (Table 16).

More than 80% of the total DOC was exported from the hillslope region to the total DOC export which is congruent with the variation of discharge and suggests a DOC flux largely controlled by the water flux term. Even though a similar late fall DOC flux was estimated for year 2 and 3, there was a significant increase in the fall DOC flux attributed to the hillslope compared to upper low gradient region DOC flux as observed at the UHS in year 3 (Table 17). Winter DOC fluxes were doubled from year 2 to year 3 for the low gradient area (UHS) when there was no noticeable increase in the DOC flux from the hillslope portion of the catchment during that same period. However, the hillslope region exhibited a response to the large winter melt event in year 3 while there was no difference in the spring melt DOC flux from the low gradient (UHS) portion of the catchment between year 2 and 3.

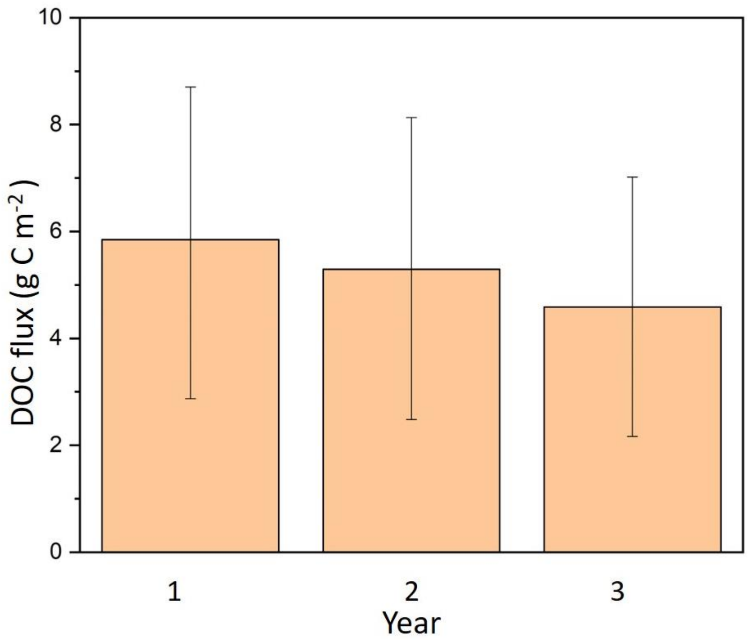

Figure 16. Lower horseshoe (LHS) annual dissolved organic carbon (DOC) fluxes in the three study years ratios. Error bars represent lowest and highest estimates of each value (1, 2, and 3 values represent year 1 (October 2014 – October 2015, year 2 (October 2016 – October 2017), and year 3 (October 2017 – October 2018)).

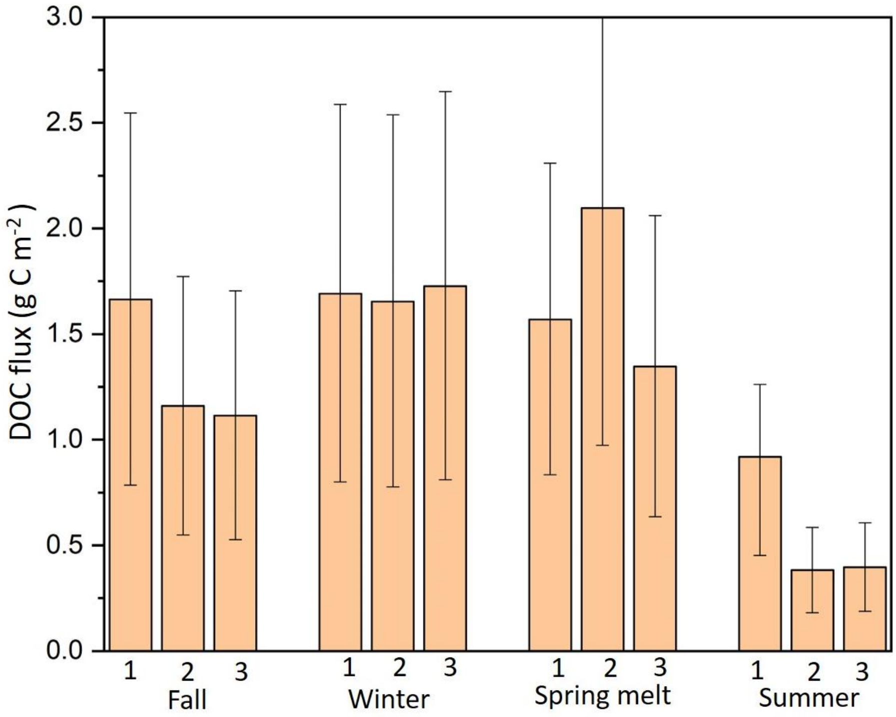

Figure 17. Seasonal Lower horseshoe (LHS) dissolved organic carbon fluxes in three years ratios. Error bars represent lowest and highest ranges of error from the actual value (1, 2, and 3 values represent year 1 (October 2014 – October 2015, year 2 (October 2016 – October 2017), and year 3 (October 2017 – October 2018)).

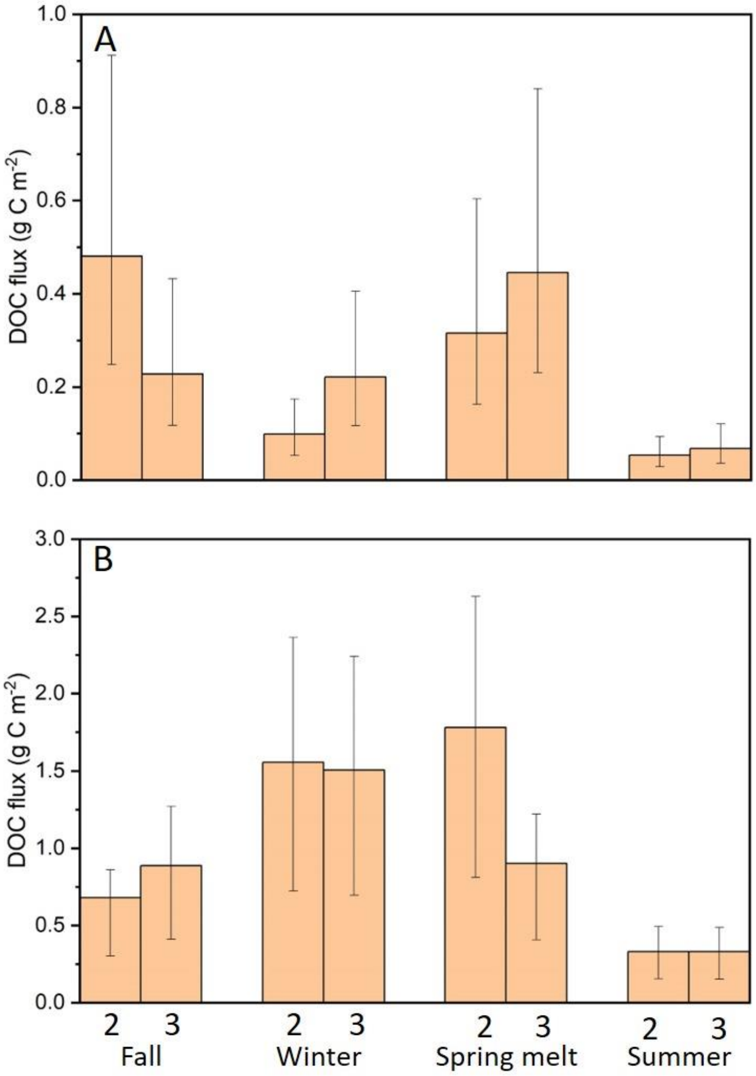

Figure 18. Upper and hillslope seasonal dissolved organic carbon (DOC) fluxes in year 2 and year 3. Ratios. Error bars represent lowest and highest estimates of each value (2, and 3 values represent year 1 (year 2 (October 2016 – October 2017), and year 3 (October 2017 – October 2018)). (A-Upper horseshoe DOC flux, B – hillslope DOC flux).

Table 16. Annual and seasonal dissolved organic carbon for the Lower Horseshoe Site (LHS). Range of lower and upper bounds of the DOC estimates are presented within brackets.

Season	Year 1 g C m^{-2}	Year 2 g C m^{-2}	Year 3 g C m^{-2}
Late fall	1.38 (0.65-2.11)	0.89 (0.42-1.35)	0.56 (0.27-0.86)
Winter	1.69 (0.80-2.59)	1.65 (0.78-2.54)	1.73 (0.81-2.65)
Spring melt	1.57 (0.83-2.31)	2.10 (0.97-3.23)	1.35 (0.64-2.06)
Summer	0.92 (0.45-1.26)	0.38 (0.18-0.59)	0.40 (0.19-0.61)
Early fall	0.28 (0.13-0.43)	0.27 (0.13-0.42)	0.55 (0.26-0.84)
Total	5.84(2.87-8.70)	5.29 (2.48-8.13)	4.58 (2.16-7.02)

Table 17. Annual and seasonal dissolved organic carbon for the Upper Horseshoe Site (UHS) and the hillslope portion of the Horseshoe Brook catchment determine from the Lower Horseshoe Site DOC minus the UHS DOC. Range of lower and upper bounds of DOC estimates are presented within brackets next to each estimate.

Season	Year 2		Year 3	
	g C m^{-2} UHS	g C m^{-2} Hillslope	g C m^{-2} UHS	g C m^{-2} Hillslope
Late fall	0.42 (0.22-0.81)	0.47 (0.20-0.55)	0.08 (0.04-0.14)	0.48 (0.22-0.72)
Winter	0.10 (0.05-0.17)	1.55 (0.72-2.36)	0.22 (0.12-0.41)	1.50 (0.69-2.24)
Spring melt	0.32 (0.16-0.60)	1.78 (0.81-2.63)	0.45 (0.23-0.84)	0.90(0.41-1.22)
Summer	0.05 (0.03-0.09)	0.33 (0.15-0.49)	0.07 (0.04-0.12)	0.33 (0.15-0.49)
Early fall	0.06 (0.03-0.11)	0.21 (0.10-0.31)	0.15 (0.07-0.29)	0.40 (0.41-1.27)
Total	0.95 (0.49-1.78)	4.34 (1.99-6.35)	0.96 (0.50-1.80)	3.62 (1.66-5.22)

3.4. Discussion

A process based understanding of variation and controls on headwater discharge is critical to predicting climate change impacts on terrestrial to aquatic water and thereby solute fluxes. This is particularly important in high latitude landscapes as they are undergoing rapid changes relevant to hydrology including a reduction in the proportion of snow as precipitation (Smithson, 2002; Devito et al., 2005; Schelker *et al.*, 2013). For example, changes in climate may reduce snow as a proportion of precipitation which can lead to reductions in stream discharge (Hamlet et al., 2005; Mote et al., 2005). This has implications for water resources of downstream regions but also signifies the potential for changes in timing and hydrological pathways which may impact solute concentration and chemistry (Raymond et al., 2016)**.** For example, trends of increasing DOC concentrations and export in higher latitude rivers over the past few decades (Boyer *et al.*, 2000; Worrall *et al.,* 2004; Raymond and Saiers, 2010) may be attributed to shifts in precipitation patterns away from snow and toward more rain in late fall through winter (Barnett et al., 2005; Huntington et al., 2016). This could be due to increased availability of DOC and its mobilization in late fall and winter relative to extraction limited conditions typical of the snowmelt period when snow is a significant portion of total precipitation (Bowering et al., 2019; Haei et al., 2013).

In this study, I demonstrate support for the hypothesis that reduced annual stream discharge is associated with a reduction in the fraction of precipitation as snow (Berghuijs et al., 2014) and likely increased mid-winter melt events. Furthermore, by separating the response of the hillslope and low relief portion of this boreal forest catchment, I suggest that much of the stream discharge is derived from runoff from the steep hillslope portion of the watershed which likely controls seasonal responses and interannual difference in overall discharge. The upper watershed does not contribute as much discharge and suggests that timing of lake ice formation and break-up along with the low gradient topography contribute to the water storage in this part of the catchment. Contrasting results were discussed in some studies that upland forested catchments such as wetland catchments contribute more flow and DOC fluxes in high latitude watersheds (Andersson and Nyberg, 2008; Devito et al., 2005; Winterdahl et al., 2014). However, findings of this study suggest that hillslope-dominated catchments might be susceptible to immediate impacts of climate change, while low gradient regions may serve to buffer watersheds against such shifts associated. However, over longer time scales wetland and pond drying or expansion might create even larger impacts in response to climate change.

3.4.1. Intra annual and inter annual variation in discharge within a small boreal forest stream

Snow cover and its variation is an important feature regulating the hydrology of boreal streams. Therefore, climate change impacts on both the magnitude and dynamics of the snowpack (Cayan et al., 2001; Stewart et al., 2005) likely represent important factors regulating boreal stream discharge. Recent climate change may alter the precipitation patterns including increases in the proportion of rain during the winter season (Altdorff et al., 2017; Berghuijs et al., 2014; Mote et al., 2005; Skjelkvåle et al., 2005; Solomon et al., 2007). Increased temperatures in winter and spring periods associated with changing climate result in increased winter runoff, earlier spring melt runoff and reduced summer streams flow (Devito et al., 2005; Raymond et al., 2016). The comparison of three different years contrasting in snowpack dynamics suggests reduced snowfall as proportion of precipitation or degradation of winter snowpack reduces annual discharge. Expectantly, the large mid-winter melt observed in year 3 resulted in a reduction in spring melt discharge and runoff. However, despite the fact that the precipitation was greatest in year 3, there was a reduction in annual discharge and runoff ratio relative to the other two study years where the snowpack was relatively consistent or experienced only a few small melt events during the winter.

Seasonal differences in the discharge observed here suggest that spring melt represents an important discharge period with hydrologic dynamics that facilitate stream flow. Soil saturation associated with extensive snowmelt can activate surface and subsurface flow paths that may efficiently connect catchment water sources to streams, thus defining many high flow events (Fiebig et al., 1990; Jennings et al., 2010; Worrall et al., 2006). My results here further suggest that these intense periods can also be important to streamflow generation on annual time scales such that precipitation or temperature regimes that reduce snowfall or snowpack contributing to the spring melt event can lead to a significant reduction in annual stream discharge. Discharge is highest during the snowmelt periods in many northern boreal catchments often with most of the solutes and organic matter being transported during these high flow periods (Hinton et al., 1997; Schiff et al., 1997). Macropore flow may lead to a greater amount of discharge during winter and spring melt periods (Noguchi et al., 2001, 1999). When macropore hydrological flow paths are activated during the high flow periods associated with these melts they can have a strong impact on solute concentration and composition in forest streams as they promote subsurface flow systems in forested soils (Tsuboyama et al., 1994; Tsukamoto and Ohta, 1988). For example, highest solute concentrations can occur during such melt periods when solutes are flushed from the upper soil horizons (Boyer et al., 1997; M. C. Eimers et al., 2008; Tsuboyama et al., 1994). This is likely because this activation from the snow water storage source occurs after a long period of drainage, a consequence of the consistent snowpack in snow dominated winters. Many studies indicate that most of the total organic carbon export occurs during spring melt period when a greater proportion of the annual discharge and runoff occur during that period (Köhler et al., 2008; Laudon et al., 2007; Rantakari et al., 2010).

Winter period melts, like observed in year 3, highlights a potential consequence of recent climate change (Jones and Perkins, 2010; Schelker et al., 2013; Varhola et al., 2010). These winter snowmelts affect stream response by influencing the hydraulic flow paths associated with hillslopes in the same manner as the spring melt (Sidle et al., 2000; Tsuboyama et al., 1994) but at different times of the year that may be relevant to solute sources from those hillslopes. The magnitude of these fluxes may change as a consequence of the timing of that intense discharge as earlier mobilization in winter may mean more organic matter available for extraction then what may be available in the following spring. The concentration and flux of soil DOC, as well as the chemical nature of the soil dissolved organic matter, in forest hillslopes within the catchment studied here suggest significant microbial processing under the snowpack over winter and thereby reduction in soil DOC sources during the spring melt period (Bowering et al., 2019). Collectively this work suggests that earlier winter snowmelt events and reduction in the magnitude of spring snowmelt has the potential to increase DOC export from forest hillslope streams.

Reduction in the development of the snowpack and the loss of the snowpack before the usual snowmelt period can reduce the spring melt, and perhaps alter solute transport, but may also reduce the overall annual discharge. In boreal forested watersheds, snow acts as a storage component of water (Hamlet et al., 2005; Mote et al., 2005; Worrall et al., 2004) which acts as a great source of water during the following spring melt period. Rapid release of that storage due to the melting events during the winter goes into recharging components and macropore filling deposits of the landscape and likely less contributing to lateral flow within the snowpack or surface organic layers in the landscape. Quick melting of snowpack particularly in hillslope regions of the catchment also likely increases direct runoff providing less time for infiltration. The reduced annual runoff ratio in year 3, where the greatest mid-winter melt occurred may suggest that a reduction in the spring melt caused by the complete snowpack loss that occurred mid-winter that year can reduce annual discharge and runoff as observed in that year relative to the other two in this study. Variation in annual runoff can also be attributed to changes in water storage in response to precipitation amount. For example, in year 2 precipitation was reduced by about 20% relative to both year 1 and the following year 3. This could have reduced water storage in the catchment and reduced the runoff in year 3, however, the year-to-year patterns in seasonal runoff ratios suggest this is likely not the case in this humid boreal landscape. The runoff ratio observed in the late fall, the first season observed in year 3, was similar to that observed in year 2 which represented a late fall period following a wetter year exhibiting the highest runoff ratios in summer and early fall. Exploring the relationship between runoff ratio and seasonal distribution of discharge, such as proportion of spring snowmelt discharge to the total annual discharge, would help to determine if reduction in annual discharge can be attributed to reductions in spring melt runoff in this mesic boreal forest landscape.

3.4.2. Variation in discharge and runoff ratios for two common and contrasting elements of boreal forest catchments.

To improve our prediction of the impacts of climate change in terms of boreal catchment water balance, it is important to understand how key landscape components (1) contribute to water inputs, and (2) respond to changes in precipitation form, distribution and quantity. Recently deglaciated landscapes of boreal regions have resulted in poorly developed drainage networks where both steep hillslope and low relief pond and wetland regions contribute to drainage into headwater stream systems. The results of the study of both the upper low gradient and lower steep hillslope elements of a boreal forest catchment indicate the impacts of climate change induced snowpack changes will likely depend on the proportion of these elements. This is likely due to the fact that topographical and morphological components have a strong influence on stream discharge and runoff ratios (Mcglynn et al., 2004; Seibert and Mcglynn, 2007; Welsch et al., 2001), and thus an interaction between these catchment elements and shifts in the form and quantity of water can result in significant changes in stream discharge responses.

Steep hillslopes in boreal regions with permeable shallow soils promote rapid lateral runoff to adjacent streams, therefore, these elements can be more responsive to changes in snowmelt reductions. In this study, the hillslope region contributes more than 80% of the stream discharge of this small stream despite making up about the same catchment area as the upper low gradient region of the catchment. The dominance of the hillslope contributions to discharge is congruent with the consistency observed in patterns of total discharge and that observed solely for the hillslope region alone. During the spring melt period, the relatively high wetness conditions promote a hydrological connection between the hillslope, wetlands, riparian and stream interfaces (Devito et al., 1996; R C Sidle et al., 2000). Pathways of subsurface flow in hillslope show a complex interaction with antecedent wetness and macro-porosity (Tsuboyama et al., 1994). Macropore flow generally occurs during the peak periods and the recession limbs of larger storm event hydrographs (Sidle et al., 1995; Tsuboyama et al., 1994) that could promote more hillslope discharge during the spring melt periods. Therefore, it makes sense that the hillslope region is highly associated with runoff generation during periods of melting. Reduction in the magnitude of that snowmelt also means the hillslope region is susceptible to large reductions in discharge and the runoff ratio in the spring melt period as observed in this study.

In contrast with the very responsive hillslope region of this catchment, the discharge and runoff ratios for the upper low gradient region remained quite comparable throughout the study period and were unresponsive to the variation in melting events among the three study years. This is somewhat surprising given that direct snowmelt inputs have been found to contribute significantly to boreal wetland, or mire, dominated headwater streams (Kohler et al. 2004; Petrone et al. 2007). Low relief portions such, however, as these can act as a storage of water in these boreal landscapes particularly given the small lakes in this landscape (REF?). Storage in the ponds and wetlands of these low relief areas contrast with the storage in hillslopes which is dependent upon the snowpack as exemplified by mountainous watersheds (Hamlet et al., 2005; Mote et al., 2005). In contrast to mountainous streams, boreal stream responses to snowmelt shifts attributed to climate change will differ because of the contributions of these low relief areas which may dampen stream responses to mid-winter melting events and the loss of snowpack. Therefore, the proportion of low relief and hillslope regions of the landscape will be helpful to predict the responses of stream discharge in many boreal catchments associated with the changes of snowpack and snowmelt dynamics.

3.4.3. Hypothesized implications of changing discharge dynamics on stream DOC as an example of impacts on terrestrial-to-aquatic solute fluxes.

Differences in discharge and runoff ratios caused by shifts in snowpack and snowmelt dynamics can have important implications for solute transport including DOC. This is because of changes in the flow path and thereby source and processing of solutes en route to the aquatic environment. Different landscape portions studied here signify important differences in both hydrology and flow paths as well as responsiveness to changes in snowpack and snowmelt dynamics. These factors suggest the potential for changes in solute transport including DOC as a consequence of these changes in the snowpack.

Annual and seasonal patterns in discharge and its sources are important in understanding the solute fluxes such as DOC. For example, annual patterns in total organic carbon flux are typically attributed to variation in stream water discharge rather than carbon concentration (Urban et al. 1989; Clair and Ehrman 1998; Hope et al. 1994; Kohler et al. 2008). The DOC fluxes in this study varied within the range of 2.16 – 8.80 g C m^{-2}, and thereby within the range of other studies but measurements used here were not frequent enough to determine variability in DOC concentration and its association with precipitation and discharge (Köhler et al., 2008; Mortsch and Quinn, 1996; Mulholland, 2003; Rantakari et al., 2010). Reported DOC fluxes in each of the three years on a seasonal basis are congruent with the variation of the discharge. DOC concentration could be increased due to the reduced proportion of hillslope water which is coming into the system as a result of reduced snowpack and snowmelt period – dilution effect often observed in other boreal streams (Kane et al., 2005; Lepistö et al., 2014). However, the magnitude of the DOC flux specific to the low relief portion of the catchment appeared to be greater relative to its discharge when compared with the entire catchment or the hillslope region. This may suggest that the upper low relief area contributes a larger portion of the DOC flux relative to the contribution of water fluxes. These findings are consistent with observations of the positive correlation between wetland contributions in catchments and DOC concentrations and fluxes in rivers including within boreal landscapes (M. C. Eimers et al., 2008; Rantakari et al., 2010; Striegl et al., 2005). In boreal landscapes, DOC concentrations and DOC fluxes are associated with the proportion of wetlands and the major sources of organic carbon substances in glaciated catchments are wetlands. Major sources of organic carbon substances in glacial catchments come from wetland regions (Hemond,1990) and DOC concentrations and DOC fluxes are also mainly associated with the proportion of wetlands (Andersson and Nyberg, 2008; Mulholland and Kuenzler, 1979; Urban et al., 1989). Reducing the inputs of water from hillslope areas may increase the proportion of inputs from low relief area to the watershed which contains more DOC coming from wetland regions. This mechanism could represent a hypothesis for the observed increased DOC in some high latitude streams over the past decades and in association with climate change.

Additionally, the estimated DOC fluxes presented here suggest that they are greatly reduced relative to DOC mobilized in the forest soils indicating high levels of processing and absorption within the terrestrial to the aquatic interface.

Bowering et al. (2019) indicated that the actual movement of DOC from the surface moss and organic soil horizon of the catchment forests is 54 - 38 g C m^{-2} on an annual basis which is more than 6 times higher in the magnitude of stream DOC flux estimates made here (Bowering et al., 2019). However, these estimates lead to the understanding that the mobilization which occurs in the soil is much larger than that was observed in the streams especially in hillslope areas suggesting that large processing of DOC occurs in forested regions. However, these estimates lead to the understanding that the mobilization which occurs in the soil is much larger than that was observed in the streams especially in hillslope areas suggesting that large processing of DOC occurs in forested regions.

More accurate stream DOC fluxes can be obtained through continuous measurements which coupled with the water fluxes as obtained here. These approaches will allow us to test the hypothesis of increases in the proportion of low relief catchment components to stream discharge as a result of reduced snowpack and snowmelt potentially due to climate change can increase DOC concentrations and potential export. Furthermore, quantifying more precise stream DOC fluxes may allow us to better constrain the timing of terrestrial to aquatic DOC fluxes and how they are linked to terrestrial fluxes. This will enable a better accounting of C fluxes at a landscape scale and also contribute to an understanding of where and when transformations of terrestrial organic matter occur and how these fluxes are controlled.

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4.0 Summary

Boreal headwater streams are important in controlling water and solute fluxes in high latitude watersheds and represent sites for the study of terrestrial to aquatic fluxes. The fact that headwater streams are highly vulnerable to recent climate change emphasizes the need for quantification of intra and interannual discharge and solute fluxes to better assess their responses and thereby the impact of climate change on important landscape-scale fluxes. However, appropriate rating curves are often difficult to obtain for the remote and rocky streams typical of boreal landscapes.

Accurate quantification of headwater stream discharge is key to understanding the effects of snow dynamics, precipitation regime, and the behaviour of landscape elements in catchment hydrology of boreal watersheds. These are all aspects needed to better predict water and solute fluxes with climate change. Results of chapter 2 showed that the methods that rely more on field data such as channel cross section, channel gradient, grain size were quite successful in predicting discharge for unmeasured gauge heights in high latitude streams especially during the high flow conditions. Among the used methods, the Ferguson resistance equation was the most successful in estimating discharge in the headwater streams as verified by comparison with regional runoff values. Relative roughness of the channel which is attributed to the grain size of the stream channel appears to be a prominent factor for predicting stream discharge as applied in the Ferguson resistance equation. Extending rating curves of high latitude streams for the full range of gauge heights using this approach helps in the effort to obtain more accurate estimates of stream discharge thereby improving our ability to capture water and solute fluxes in boreal landscapes.

There is a clear need for further research studies in identifying the factors that affect the flow resistance of these high latitude boulder bed streams and applying those factors in flow resistant equations will provide more accurate estimates of stream discharge. Increasing the number of field discharge measurements for each stream site will expand the range of data which could be more useful to add more accuracy to discharge estimates. Furthermore, there should be more steps taken to develop methods of assessing the uncertainty of these rating curves. Several studies assessed the uncertainty around the stream discharge estimates for larger river systems (Clark et al., 2015; Coz, 2012; Domeneghetti et al., 2012), but it is quite important to expand those methods of estimating the uncertainty of stream discharge for small streams. Increasing the accuracy of the measurements at the field and the replicating the number of measurements will reduce the variability of the predicted discharge. A better estimation of the actual uncertainty of the water flux measurements based on rating curve development allows for more interpretable stream discharge and also it can be quite useful in future studies for an accurate understanding of annual and seasonal variations of water fluxes, as well as the solute fluxes derived from water fluxes, within these catchments.

Snowmelt represents an important period for stream discharge in boreal landscapes (Schelker et al., 2013; Schelker et al., 2013; Huntington et al., 2016;

McCabe et al., 2018). With changes in the timing, form and magnitude of winter precipitation and consequential spring snowmelt dynamics as a consequence of climate change (Barnett et al., 2005; Hamlet et al., 2005; Mote et al., 2005; Stewart et al., 2005), we expect significant alterations in the headwater stream discharge dynamics. For example, reductions in the stream discharge were experienced with the significant reductions in the winter snowpack in many snow dominated watersheds (Berghuijs et al., 2014). Decreased spring melt discharge in year 3 suggested that mid-winter melts reduced the following spring melt discharge and likely annual discharge overall, therefore snowpack dynamics and the winter precipitation regime act as key factors of controlling stream hydrology within these watersheds.

The study of low relief (wetland and pond dominated area) and hillslope components of this headwater catchment indicated that the majority of water (80%) was attributed to the hillslope regions despite representing about 50% of the total catchment area. Furthermore, the hillslope region was responsive to the mid-winter melting events whereas no significant changes were observed in water fluxes for the low relief area as a consequence of the different snowpack dynamics captured in this study. The reduction in overall hillslope discharge as a consequence of mid-winter snowpack loss and decreased snowmelt discharge suggests that climate change will likely enhance the proportion of waters draining low relief wetland areas relative to hillslope regions. If realized across boreal forest streams, this indicates great potential for shifts in solute composition, concentration and perhaps even flux with climate change. For example, this finding provides evidence for a potential mechanism for climate change driven increases in DOC concentration. River and stream DOC concentration and flux are typically associated with wetland coverage in the catchments of boreal and other landscapes (Huntington et al., 2016; Rantakari et al., 2010; Worrall et al., 2004). If hillslope contributions to streamflow are reduced, as suggested here, with reductions in snowpack and snowmelt events, increased proportion of low relief contributions to flow will bring with it increases in DOC concentration.

Given the high level of uncertainty on DOC estimates, obtaining more accurate DOC fluxes have become more challenging especially for remote headwater streams. However, more continuous measurements of stream DOC concentrations (e.g. in situ optical instrumentation), including four seasons particularly for different hydrological events, would be more helpful in addressing these challenges. Obtaining DOC estimates for contrasting years of precipitation would also be important to explore the mechanisms of controlling stream DOC with respect to changing the climate. Furthermore, the knowledge of these mechanisms will enhance our understanding of controls on the terrestrial to aquatic export of DOC in these snow dominated forested watersheds. Overall, estimates of stream DOC fluxes for this catchment suggested that these changes in precipitation regime and proportion of catchment components may affect both the composition and fate of DOC in the aquatic environments.

5.0 Reference

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