Streamflow characteristics of intermittent streams in the Okanagan, British Columbia, Canada

by

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B.Sc. (Hons.), Simon Fraser University, 2006

Thesis Submitted in Partial Fulfillment of the
Requirements for the Degree of
Master of Science

in the
Department of Geography
Faculty of Environment

SIMON FRASER UNIVERSITY
Fall 2018

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Date Defended:	August 24, 2018

Abstract

Even though intermittent streams are prevalent in every hydrologic region of the world, there has been relatively little research on the flow regime of intermittent streams. Intermittent stream flows affect aquatic biota, water quality, and quantity in downstream perennial streams. This study focused on (1) the differences in the streamflow regime characteristics of perennial and intermittent streams and (2) the spatial and temporal variation in streamflow along one intermittent stream (Long Joe Creek) in the Okanagan Basin. Differences between intermittent and perennial streamflow regimes included rates of recession, steepness of flow duration curves, start of freshet, and fall discharge variability. Lack of streamflow data for truly intermittent streams made it difficult to assess the reasons for these differences. The observations along Long Joe creek highlighted the very large spatial and temporal variability in streamflow in intermittent streams and the need to focus monitoring on multiple locations.

Keywords: intermittent streams; Okanagan; flow regimes; spatial variation; gaining and losing reaches

Dedication

To Ilja, Shayne, Mum and Dad. You never stopped believing that I would finish. Each one provided an enormous amount of time and energy to help me compete this thesis. That support kept me going.

Acknowledgements

The journey to complete this thesis was not short and many life milestones were passed; including the birth of two children. This document would not be here today without support and encouragement from numerous wonderful individuals – my thanks to all of them for being a part of this journey and making this thesis possible. I owe my deepest gratitude to Dr. H.J. (Ilja) van Meeveld. Not only did Ilja supervise this project from start to finish, she provided never ending patience, support, edits, and knowledge. Dr. van Meerveld always found time to answer my questions, provide feedback and give me guidance – for that I thank you.

I would like to express my gratitude to Dr. P.H. Whitfield, who inspired me to go back to school and piqued my interest in intermittent stream hydrology. I am grateful to Dr. Leach who stepped in as a Supervisor and provided encouragement. I also want to express my warmest thanks to Dr. Lesack. He made the final push happen by offering to take on Supervisory duties, organizing the defence and providing support especially towards the completion of the thesis. I am thankful to Environment and Climate Change Canada for the equipment and data used during the study and for the time provided, without these the thesis would not have been possible.

Many thanks also to the Hydro Lab crew who always made coming to campus and working in the lab a joy. I specially loved the chocolate eating, and dinner parties.

Finally, my deep and sincere gratitude to Mum, Dad and Shayne for their continuous and unparalleled love, help and support. Shayne was the cornerstone of this thesis, I am forever indebted to him for all that he has given to make this happen.

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Preface Image

VIEW ON LONG JOE CREEK NEAR OSOYOOS BRITISH COLUMBIA – Feb. 20, 2011

Chapter 1. **Introduction**

Arid and semi-arid ecosystems cover about one third of the earth's land surface (Whitford, 2002). These regions are characterized by low and highly variable precipitation, so that rivers and streams are often ephemeral or intermittent. Ephemeral streams or stream reaches are located above the groundwater table and flow only in direct response to precipitation. Intermittent streams or stream reaches, which are the focus of this study, flow only at certain times of the year, for example in response to periods of snowmelt or precipitation. Losses along the stream channel can be high for intermittent streams, so that flow at the outlet is only recorded when the flow from upslope and upstream areas is larger than the transmission losses (Shanafield and Cook 2014; Prudic et al. 2003; Lange 2005)

Intermittent streams are also common in more humid environments. It has been estimated that intermittent streams constitute more than 50% of the global river network (Datry et al., 2014; Larned et al., 2010). Intermittent streams account for 69% of the first order streams (below 60° latitude) and 34% of the fifth order rivers globally (Raymond et al. 2013). Their prevalence is expected to increase in the future due to drying in response to climate change and increased water extractions (Larned et al. 2010). An overview of temporary streams in Canada by Buttle et al. (2012), concluded there is a lack of studies on Canadian temporary streams and rivers.

Hydrological processes in catchments with intermittent streams have received less research attention than for processes in perennial catchments and remain poorly understood (Raymond et al. 2013; Larned et al. 2010; Datry et al. 2014). There are very few studies that describe how to measure and determine flow in intermittent streams (Peters et al. 2012; Buttle et al. 2012). Streamflow in intermittent streams cannot be predicted by extrapolation from perennial streams (Scott 2006) because their response to variations in precipitation tend to be highly non-linear. Flow in intermittent streams and rivers is governed by irregular flow patterns and streamflow is highly variable in both space and time (Bull and Kirkby 2002). This makes intermittent streams difficult to monitor. Because sediment transport can be high for intermittent streams and flows can be very low, standard gauging techniques are often not accurate enough. There are also no standards for how to report zero flow (Peters et al. 2012). In Canada, very few intermittent streams are monitored continuously, despite their prevalence across the country (Peters et al. 2012).

Few studies have focused on how streamflow varies along stream networks, even though this provides useful information on the balance of inflows and outflows. The spatial variability in streamflow is particularly pronounced for intermittent streams because sections of the stream go dry, while other sections can continue to flow during the dry season. Understanding and predicting patterns of intermittent streamflow in time and space and the associated ecological consequence of altering patterns of flow variability are fundamental to water resources management (Larned et al. 2010; Levick et al. 2008) because intermittent streams provide groundwater recharge (Shentsis and Rosenthal 2003; Izbicki 2007; Rushton 1997), transport sediment and nutrients (Martin-Vide et al., 1999; Scott, 2006), and are critical habitat for many species (Gomi et al. 2002; Datry et al. 2014; Dodds et al. 2004). The expansion and contraction of streamflow in intermittent streams is one of the most important determinants of ecological patterns and processes in rivers (Power et al. 1995; Richter et al. 2003). Knowledge of flow processes in intermittent streams is also essential for understanding and protecting ecosystems in downstream perennial streams because they are directly linked.

In arid and semi-arid environments intermittent streams can be a significant source of water for downstream rivers and lakes. This is likely also the case for the Okanagan in British Columbia Canada, where intermittent streams are common but few intermittent streams are gauged. The Okanagan Basin Water Board, created in 1969 to address concerns about water allocation, quantity and quality in the Okanagan Basin, oversaw the Water Supply and Demand Project to assess water supply in the Okanagan basin. This assessment concluded that there is a lack of information about streamflow-groundwater interactions and understanding of the flow regimes of small streams during low and no flow periods (Guy 2010).

Therefore, this thesis focuses on intermittent streams in the Okanagan. In chapter 2 , the five components of a streamflow regime (magnitude, frequency, duration, timing, and rate of change) are used to explore the streamflow characteristics of intermittent streams and to compare them to perennial streams. This chapter was published in the special issue of Canadian Water Resources Journal on "Prediction in Ungauged Basins (PUB)

Workshop on Temporary Streams". Chapter 3 is a detailed study of a small catchment with an intermittent stream in the Southern Okanagan and focuses on the occurrence and distribution of streamflow and how this is related to topographic and catchment characteristics. It also evaluates the usefulness of surface temperature measurements to determine the spatial and temporal patterns in the occurrence of flow along the stream. Chapter 4 summarizes the main results from the two studies and provides suggestions for future work.

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Chapter 2. **Intermittent and perennial streamflow regime characteristics in the Okanagan**[1]

2.1. Abstract

Streamflow data from ten Water Survey of Canada gauging stations were analyzed to characterize streamflow regimes in the Okanagan Basin (British Columbia). The differences in the streamflow regime characteristics of the perennial and intermittent streams were subtle, except for the obvious difference in summer low flows. The intermittent streams tended to have faster recessions after spring freshet, steeper flow duration curves, a slightly earlier median day of the year of the start of the freshet, and more variable discharge in fall. In years with high fall precipitation, discharge was high in fall for the intermittent streams but in other years it was very low. Discharge on August 15th was lower or similar to streamflow on March 15th for the intermittent stream, whereas it was much higher on August 15th for the perennial streams. These subtle streamflow regime differences point to differences in flow pathways, groundwater contributions to streamflow, and residence times between the intermittent and perennial watersheds, and may have important ecological implications.

Keywords: Okanagan Basin, streamflow regime, intermittent stream, watershed characterization

[1] This chapter has been published as:
Huxter, E.H., and H.J. van Meerveld (2012) Intermittent and Perennial Streamflow Regime Characteristics in the Okanagan, Canadian Water Resources Journal / Revue canadienne des ressources hydriques, 37:4, 391-414, DOI: 10.4296/cwrj2012-910

2.2. Introduction

Streamflow regimes describe the characteristics of streamflow quantity, timing and variability (Poff *et al*., 1997). They influence channel geomorphology and the distribution and abundance of riverine species (Resh *et al*., 1988; Power *et al*., 1995). Streamflow regimes have a profound influence on the biodiversity of riverine ecosystems (Poff *et al*., 1997; Richter *et al*., 1997; Hart and Finelli, 1999, Poff and Zimmerman, 2010) because hydrologic variability is important for riverine ecosystem health, structure, and function (Poff and Ward, 1989; Datry and Larned, 2008; Poff *et al*., 2009). Natural hydrologic variation is integral to biotic diversity within aquatic ecosystems (Arthington and Pusey, 2003; Richter *et al*., 1996; Lloyd *et al*., 2003; Poff and Zimmerman, 2010). Alterations of natural flow regimes affect the entire riverine ecosystem. For example, modification of the timing, duration or magnitude of floods can eliminate spawning cues for fish (Junk *et al*., 1989; King and Louw, 1998). An increase in the frequency, duration or rate of change of high streamflow can displace velocity sensitive organisms such as phytoplankton and macroinvertebrates (Allan, 1995), each then affect other biotic assemblages in the ecosystem. This has led to a large body of scientific literature advocating the use of natural streamflow regimes as a guide for ecosystem and water resource management (e.g., Poff *et al*., 1997, Richter *et al*., 1997, Stromberg, 2001; Nilsson and Svedmark, 2002; Hauer and Lorang, 2004; Poff *et al*., 2009).

When water resources become scarcer and demand increases, wise water resource management is needed to prevent damage to riverine ecosystems. Billions of dollars are spent annually to restore rivers to their “natural state” (Palmer *et al*., 2004; Bernhardt *et al*., 2005). Understanding and predicting natural patterns of streamflow in time and space and the associated ecological consequences of altering these patterns of flow variability have thus become fundamental to water resource management (Bunn and Arthington, 2002; Arthington and Pusey, 2003; Richter *et al*., 2006; Kennard *et al*., 2010). However, many years of observation are generally needed to describe the natural flow regime (Chang *et al*., 2011).

Analysis of the five streamflow regime components (magnitude, frequency, duration, timing, and rate of change of hydrologic conditions) that regulate ecological and physical processes in river ecosystems (Poff and Ward, 1989; Walker *et al*., 1995; Richter *et al*., 1996; Poff *et al*., 1997; Greet *et al*., 2011), allows for the characterization of streamflow responses and the hydrologic behaviour of watersheds.

- **Magnitude** refers to the discharge and is the amount of water moving past a given point in a given unit of time. Magnitude can be relative or absolute depending on the nature of the research question. Climate, geomorphology and watershed size have a large impact on magnitude.

- **Frequency** describes how often a particular discharge event occurs in a specified period of time. Frequency can be analysed at many different time scales: events based on storms, freshet, or decadal cycles.

- **Duration** is the period of time associated with a specific flow event. Duration can be relative to a particular flow event (days of flow in a temporary stream) or it can be expressed as a composite, over a specified period of time (e.g. days that flow exceeds the annual mean/maximum/median).

- **Timing** or predictability of flows of a specified magnitude is the regularity with which a certain streamflow occurs. This can be defined formally or informally and with reference to different time scales. For example, annual minimum flows may occur with low seasonal predictability or with high seasonal predictability.

- **Rate of change** refers to how quickly discharge changes in magnitude in any given event or seasonally. If a river is stable there is a very slow rate of change, whereas for flashy rivers discharge changes rapidly.

Streamflow regime analysis provides not only information on flow variability, the magnitude of peak and low flows, and the duration of zero flows, but also on the principle transport frequency of sediment and dissolved load, and possible groundwater or natural storage influences. This information is important for fluvial geomorphologists and hydrologist and has practical value to stream ecologists (Reece and Richardson, 1995; Poff *et al*., 2009). Knowledge about streamflow regimes can also aid in the prediction of streamflow in ungauged basins as streamflow regimes are signatures of the interactions between climate and watershed characteristics. Streamflow regimes for ungauged watersheds can be inferred based on surrounding stations with long-term streamflow records or constructed using geo-statistical or modeling approaches based on the characterization of streamflow regimes in similar (or neighbouring) watersheds (Kennard *et al*., 2010; Sauquet and Catalogne, 2011). Moore *et al*. (2011), for example, used a spatially distributed water balance model to predict inter-catchment variation in annual runoff and the seasonal distribution of runoff in British Columbia (BC). The model was able to distinguish between pluvial, nival and hybrid streamflow regimes and predicted annual runoff with varying levels of accuracy. Sandborn and Bledsoe (2006) used multiple regression models based on data from gauged watersheds and physical and climatic characteristics to predict streamflow regime characteristics in ungauged watersheds in Colorado, Washington and Oregon. Their multiple regression models could predict the magnitude, timing, and rate of change variables quite well but were less successful in predicting streamflow variability. Similarly, Viola *et al*., (2011) and Saquet and Catalogne (2011) used regression equations to derive flow duration curves for ungauged watersheds in Sicily and France, respectively.

Because very few intermittent streams are monitored for extended periods, little is known about the streamflow regimes of intermittent streams in Canada and how they differ from the flow regimes in neighbouring perennial streams, except for the obvious difference that there are periods of zero flow. This paper presents an overview of the broad characteristics of streamflow regimes in the Okanagan Basin in British Columbia (BC) using streamflow data from ten Water Survey of Canada (WSC) stations. Specifically, we look at the *i)* hydrographs and streamflow variability, *ii)*, spring freshet, *iii)* summer low flows, *iv)* fall peakflows, and *v)* cumulative frequency distributions (flow duration curves), and highlight similarities and differences in the magnitude, frequency, duration, timing and rate of change for intermittent and perennial streams. By comparing streamflow regimes of watersheds within a similar climatic region and correlating the streamflow regime characteristics with basic physical watershed characteristics, we aim to improve our understanding of the behaviour of these watersheds and the controls on particular streamflow regime characteristics within the Okanagan Basin. This study is, to our knowledge, the first attempt to characterize perennial and intermittent streams in the Okanagan Basin. These initial results could later be used for regionalization of streamflow regimes to aid prediction in ungauged watersheds in the Okanagan Basin.

2.3. Site description

The Okanagan Basin is located in southern BC (Figure 2.1). The valley is 5-10 km wide in the south near Osoyoos, and as wide as 18 km in the north around Armstrong, with a total basin area of approximately 8000 km^2. The valley bottom ranges from 416 m above sea level (m asl) near Vernon to as low as 270 m asl in Osoyoos, while the upland alpine tundra is as high as 2400 m asl. Lakes cover much of the valley floor and are fed by perennial and intermittent streams on the surrounding hillslopes. Most of the streams feeding the main valley are in narrow, deeply incised valleys and have steep gradients. These surrounding mountain streams deliver the majority of the water to the Okanagan Basin during the spring freshet. Intermittent streams in the region are understudied but could provide a significant amount of water to the basin.

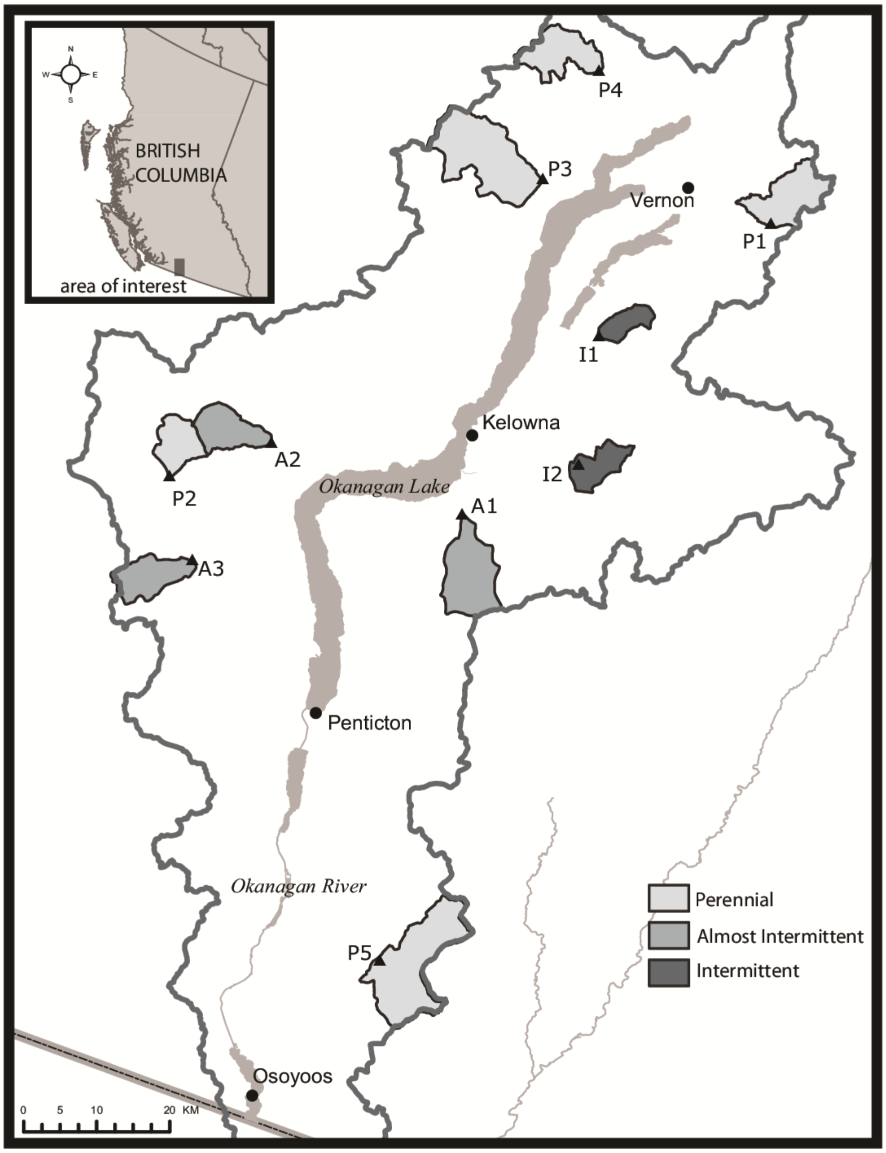

Figure 2.1 **Location of the study watersheds. Dark shading represents the intermittent watersheds, medium gray shading represents almost intermittent watersheds and light gray shading represents the perennial watersheds. The boundaries of the Okanagan Basin, Okanagan Lake and other lakes are shown as well.**

A classification of hydrologically similar regions in British Columbia using a five-parameter model for flow duration curves by LeBoutillier and Waylen (1993) showed that compared to other parts of BC, perennial streams in the Okanagan have a low annual discharge due to low annual precipitation and that the annual hydrographs are characterized by spring snowmelt followed by storage depletion. Eaton and Moore (2010) surveyed the regional hydrology of British Columbia by examining seasonal streamflow regimes and peak flow characteristics of perennial streams and characterized the Okanagan Valley by low mean annual streamflow, annual peaks generated by snowmelt (May-June), monthly streamflow values above 25% of the mean annual flow, and large year to year variations in annual average streamflow.

The climate of the Okanagan is arid to semiarid. Annual average precipitation in the valley floor ranges from 300 mm in Osoyoos to 409 mm in Vernon (Environment Canada, 2011a), while the sub-alpine area receives approximately 1200 mm (Government of British Columbia Ministry of Transport, 2011). Precipitation is fairly evenly distributed throughout the year, although May, June, November and December are slightly wetter than average. The Okanagan basin lies in the rain-shadow of the coastal mountains, with the western side of the basin receiving the least amount of precipitation. The annual average valley temperature is 8 ^{o}C in the north and 10 ^{o}C in the south, with winter (December-January) lows below 0 ^{o}C and summer (July-August) highs in the upper twenty degrees Celsius, although it is not uncommon to have consecutive days above 30 ^{o}C in the summer (Environment Canada, 2011a).

Vegetation in the valley bottom is primarily sagebrush and bunchgrass, which become mixed with open Ponderosa pine (*Pinus ponderosa* C. Lawson) grasslands just above the valley bottom. The upper valley slopes are dominated by a montane forest of Douglas-fir (*Pseudotsuga menziesii* Mirb.) mixed with lodgepole pine (*Pinus contorta var. latifolia*). Montane spruce (*Picea Glauca* Moench) forest with alpine tundra occur near higher mountain summits (Heinrichs et al. 2001). These vegetation zones are not only controlled by elevation but also by climate, soils, aspect, and disturbance history (Heinrichs et al. 2001). Land-use in the lower elevations outside the urban areas is primarily grazing, forage production, orchards, vineyards, water-oriented recreation, and residential development. The Okanagan has experienced dramatic population growth and expansion of irrigated areas, such that water supplies are now almost fully allocated (Neilsen *et al*., 2006; Wassenaar *et al*., 2011; Okanagan Basin Water Board, 2011). Climate change is projected to reduce future water availability (Merritt *et al*., 2006; Neilsen *et al*., 2006). Current surface storage systems are expected to be unable to meet municipal and instream flow needs during “normal” precipitation years by the 2050s (Harma *et al.*, 2011).

The bedrock is composed of numerous rock types, formations, and development sequences. The east side of the valley from Osoyoos to Vernon is primarily Mesozoic and early Tertiary granitic and granitic gneissic rocks, while the west side of the Okanagan valley is mainly dominated by late Paleozoic through Tertiary granodioritic intrusive rocks and volcanic rocks.

North of Vernon the geology is predominantly Tertiary fine clastic sedimentary bedrock (British Columbia Integrated Land Management Bureau, 2011). The valley bottom is a complex network of faults that separates these very different geologies and is known as the Okanagan fault. The surficial geology is primarily the result of erosion and deposition during the last glaciation, approximately 10,000 to 15,000 years ago. Surficial material is comprised of unconsolidated glacial deposits, including glacial-lacustrine, glacial-fluvial, and ice contact deposits (Bowen *et al*., 2005).

2.4. Methods

2.4.1. Data sources

Published historical streamflow records from the Water Survey of Canada (WSC) were assessed to determine streamflow regimes within the Okanagan Basin (Environment Canada, 2011b). Land use and topographic data for the selected watersheds were derived from the Government of British Columbia Land and Resource Data Warehouse (http://www.lrdw.ca). Climatic data were derived from ClimateWNA v4.60 (Wang *et al*., 2012), a standalone application that extracts and downscales PRISM (Parameter-elevation Regressions on Independent Slopes Model) (Daly *et al*., 2002) and ANUSPLINE (Hutchinson, 1989) average monthly data for the 1961-1990 period to point data and calculates seasonal and annual climate variables for specific locations based on latitude, longitude and elevation. Each watershed was divided into 250m^2 cells and point values were determined for each of these grid cells. The mean temperature and precipitation for each watershed was calculated by averaging all grid cell values.

2.4.2. Streamflow gauging station selection

The criteria for gauging station selection required that stations had a continuous operating schedule, ten or more years of streamflow data, were unregulated, and located in the Okanagan Basin. This resulted in 15 potential stations. These stations were screened for overlapping records and a subset of 12 stations was selected. The final group of stations for analysis

Table 2.1 **Overview of the watersheds used in this study. I1 and I2 are classified as intermittent streams, A1-A3 as almost-intermittent streams and P1-P5 as perennial streams.**

Stream Watershed	WSC Station Name	Drainage area (km^2)	Years of record	Average slope (º)	Minimum elevation (m asl)	Maximum elevation (m asl)	Length of streams (km)	Drainage density (km/ km^2)	Forest* (% area)	Geology (% area)			
										intrusive**	metamorphic	volcanic	sedimentary
I1	Clark Creek near Winfield	15	15	7.0	1027	1626	17	1.13	57	0	12	88	0
I2	Daves Creek near Rutland	31	22	6.7	838	1670	46	1.48	38	0	85	15	0
A1	Bellevue Creek near Okanagan Mission	73	29	11.6	597	2163	75	1.02	83	43	57	0	0
A2	Greata Creek near the mouth	41	39	12.9	917	1703	63	1.54	92	91	0	9	0
A3	Bull Creek near Crump	47	22	11.7	916	1943	65	1.38	86	100	0	0	0
P1	Coldstream Creek above Municipal intake	59	42	13.5	610	1663	98	1.66	88	5	0	23	72
P2	Camp Creek at Mouth near Thirsk	34	44	13.1	972	1928	60	1.76	67	100	0	0	0
P3	Whiteman Creek above Bouleau Creek	112	38	14.7	640	2040	216	1.93	89	30	0	60	10
P4	Ewer Creek near the Mouth	53	16	14.4	660	1748	148	2.79	75	0	0	74	26
P5	Vaseux Creek above Solco Creek	117	39	9.1	1200	2300	176	1.50	68	70	23	7	0

* area classified as young forest, old forest, or selectively logged forest (collected in 1993)

**intrusive includes granitic, granodioritic intrusive rocks and syenitic to monzonitic intrusive rocks contained only 10 stations because two of the 12 stations had to be excluded due to missing metadata. Of the ten stations, 5 are perennial (labeled P1-P5), 3 almost-intermittent (labeled A1-A3), and 2 intermittent (labeled I1-I2) (Figure 2.1 and Table 2.1). The intermittent streams had periods of zero flow, whereas the almost-intermittent streams had minimum daily streamflow less than 0.005 m^3/s. The small number of streams limited the analyses and prevented us from doing statistical analyses to compare the streamflow regimes of the intermittent, almost-intermittent and perennial streams but did allow for qualitative comparisons of the three streamflow regimes. Furthermore, it has to be noted that the streams that were classified as intermittent streams are not characteristic of the truly intermittent streams in the Okanagan as streamflow at the intermittent stations did not cease every year. The truly intermittent streams only had one to two years of record or were not monitored continuously. P5 (Vaseux Creek) becomes intermittent approximately 14 km downstream from the gauging station but is not classified as intermittent or almost-intermittent at the station location.

Streamflow data from the 10 stations was analyzed for the January 1, 1972 to December 31, 1982 period because this was the only period that all stations had full streamflow records. While this short period limits the analyses and the interpretation of the results, it does include wet and dry years. Based on 18 provincial government snow survey stations in the Okanagan (Government of British Columbia Ministry of Environment, 2011), 1974 was a year with an above average (+160%) snowpack, while 1981 was a year with a below average snowpack (-50%). The average of the annual streamflow as a ratio of the long term mean annual streamflow for the ten stations was 1.63 in 1974, 1.04 in 1981, and as low as 0.41 in 1973.

Mean annual precipitation for the ten watersheds during the 1972-1982 period ranged from 562 mm to 724 mm (average 636 mm, σ = 58 mm; Table 2.2). Mean annual precipitation was highest for I1 (705 mm) and A1 (724), while it was lowest for A3 (562 mm) and A2 (563 mm). Mean annual temperature for all ten watersheds was 2.9ºC (σ = 0.5ºC).

The smallest watershed is 15 km^2 and the largest watershed 117 km^2 (Table 2.1). Five of the watersheds are smaller than 50 km^2, and all but two are smaller than 80 km^2. The intermittent streams have a drainage area of 15.3 km^2 (I1) and 31.1 km^2 (I2), the smallest of all streams. However, the drainage area of the smallest perennial stream (P2) (33.9 km^2) is comparable to the intermittent streams and the almost-intermittent streams are comparable in size to the perennial streams (Table 2.1). The elevations of the gauging stations range from 597 to 1200 m asl. Minimum watershed elevation range is 600 m asl, with the lowest elevations around

600 m asl (A1, P1, P3) and the highest minimum elevation at 1200 m asl (P5). All other watersheds have a minimum elevation around a 1000 m asl. There is a small spread in the maximum elevations; most watersheds have a maximum elevation between 1600 and 1900 m asl, with the exception of A1, P3, and P5, which all extend above 2000 m asl (Table 2.1). Watershed A1 has the largest difference in elevation (1566 m), while watersheds I1 and I2 have the smallest differences in elevation (599 and 832 m respectively; Table 2.1). The watersheds range in mean slopes from 6.7^{o}C to 14.7^{o}C, with the intermittent watersheds having the lowest average slopes (Table 2.1).

Watersheds I1, I2 and P4 are the only watersheds that do not contain intrusive bedrock but are instead composed of metamorphic and volcanic rock; watershed I1 is predominantly volcanic, whereas I2 is predominantly metamorphic. Sedimentary bedrock is only found in watersheds P1, P3 and P4 (Table 2.1).

2.4.3. Data analysis

All streamflow data were normalized by drainage area in order to allow comparison of the discharge characteristics at each station. For each calendar year and each stream, the following characteristics were determined: annual discharge, the start of the spring freshet, the peak freshet, the minimum discharge in summer, and the maximum discharge in fall. Box plots showing the 10, 25, 50 (median), 75, 90th percentiles, as well as the outliers were used to represent these hydrograph characteristics because box plots show both the inter-site variability and the intra-site (year to year) variability. The start of the spring freshet was determined as the first of at least two consecutive days after March 21st that discharge increased by 0.01 mm/day and 10% from the previous day. The minimum summer flow was determined as the minimum discharge between peak freshet and October 15th. Fall peak flow was determined as the maximum discharge that occurred between the minimum summer flow and December 31st.

Based on the 11 years of data for each stream, the 10, 25, 50, 75, and 90th percentile of discharge on a given calendar day were calculated in order to determine the relative variability in streamflow throughout the year. The coefficient of variation of discharge (i.e. the standard deviation of discharge on a given calendar day divided by the average discharge on that calendar day) was determined for each calendar day and each stream as well.

Table 2.2 Overview of watershed average mean annual precipitation and streamflow characteristics for the 1972-1982 period at the stations used in this study. The mean annual discharge for the period of record is given in parenthesis for comparison. For station information, see Table 2.1. DoY=Day of Year.

	% of Zero Flow days (1972-82)	Mean annual precipitation (P) (mm)	Mean annual flow (Q) (mm)	Mean runoff Ratio (Q/P)	Median peak freshet (mm/d)	Median summer low flow (mm/d)	Median fall peak flow (mm/d)	Median DoY of start of freshet	Median DoY of peak freshet	Median DoY of summer low flow	Median Mar 15 flow (mm/d)	Median August 15 flow (mm/d)	FDC_{slope}	Master recession constant (k)
I1	0.85	705	172 (162)	0.24	5.61	0.006	0.141	94	135	239	0.034	0.034	5.1	0.911
I2	0.17	644	134 (118)	0.21	3.72	0.019	0.433	92	127	238	0.058	0.053	4.0	0.920
A1	0*	724	166 (168)	0.23	5.54	0.006	0.113	102	145	244	0.028	0.011	4.0	0.888
A2	0	563	73 (62)	0.13	1.24	0.045	0.106	101	136	241	0.051	0.079	1.9	0.958
A3	0	562	106 (92)	0.19	2.45	0.050	0.122	107	147	256	0.041	0.125	2.6	0.945
P1	0	634	137 (136)	0.22	2.88	0.071	0.130	100	126	273	0.075	0.117	3.3	0.967
P2	0	567	151 (139)	0.27	3.47	0.122	0.173	103	141	273	0.122	0.189	1.9	0.961
P3	0	647	192 (180)	0.30	5.15	0.050	0.155	97	139	260	0.072	0.096	3.0	0.936
P4	0	637	209 (219)	0.33	4.78	0.079	0.267	97	136	256	0.106	0.124	2.7	0.958
P5	0	681	245 (243)	0.36	9.75	0.086	0.427	97	145	262	0.065	0.126	2.9	0.946

*0.52% of the time streamflow was 0.001 m^3/s

Flow duration curves (FDC), showing the fraction of time that a certain discharge was equaled or exceeded, were also determined for each station. FDCs are commonly used to compare watersheds and streamflow regimes, since they display the full range of flows, including high and low flows (Vogel and Fennessey, 1995; Smakhtin, 2001). In order to objectively characterize streamflow variability, the slope between the 33rd and 66th percentiles of the FDC was calculated:

$$Slope_{FDC} = \frac{\ln(Q_{33\%}) - \ln(Q_{66\%})}{(0.66 - 0.33)}$$

where $Q_{33\%}$ and $Q_{66\%}$ are the discharges (normalized by area) that are equaled or exceeded 33% and 66% of the time, respectively. On a semi-log plot this portion of the curve is often relatively straight (Yadav *et al*., 2007; Zhang *et al*., 2008; Sawicz *et al*., 2011). Lower $Slope_{FDC}$ values represent damped responses and often indicate higher groundwater contributions and/or sustained year-round rainfall (Searcy, 1959; Sawicz *et al*., 2011).

Streamflow recession analysis provides information of groundwater discharge and watershed storage capacity (Tallaksen, 1995). Master recession curves (MRC) were therefore developed for each stream using the United States Geological Service RECESS program (Rutledge, 1998). Fifteen days of recession were required for RECESS to detect a recession period, each period was then plotted and any departures from linearity on a semi-log plot in the first few days of the recession were removed to ensure only recession data were analysed. After the analysis of the recessions, index values were calculated, outliers were removed if necessary, and the final MRC was plotted and fitted with an exponential function to determine the master recession constant (k, the slope of the line on a semi-log plot).

In order to obtain more information on the factors that affect the different flow regime components, physical watershed characteristics for which we had data (Table 2.1) were correlated with the streamflow regime characteristics using non-parametric Spearman rank correlation. Finally, the watersheds were grouped using cluster analysis. Cluster analysis is a descriptive, exploratory technique that classifies objects (i.e. watersheds) into groups based on their characteristics. The result of a cluster analysis is a dendrogram, which is a hierarchical tree that displays the distances between the objects. Cluster analysis was performed based on the watershed characteristics (Table 2.1) and streamflow characteristics (Table 2.2). In order to give all characteristics the same weight, the characteristics were scaled so that the lowest value was 0 and the highest value 1. The clusters were compared with the a priori division of the watersheds into the intermittent, almost-intermittent and perennial stream categories.

2.5. Results

2.5.1. Hydrographs and variability

The median discharge and especially the 75^{th} and 90^{th} percentiles of discharge revealed two contrasting flow regimes (Figure 2.2). The intermittent streams exhibited a flow regime with peaks during freshet, early July, and late summer/fall, which was also clear for one of the almost-intermittent streams (A1) (Figure 2.2). In contrast, the perennial streams had much more uniform flows, especially in the late summer and fall with an even and slow rate of recession. The only exception was P5, which is intermittent downstream of the gauging station, and displays slightly elevated flows in the late summer/fall (Figure 2.2). Almost-intermittent streams A2 and A3 tended to behave similar to the perennial streams (Figure 2.2).

The coefficient of variation of discharge on each calendar day, also demonstrated that flow variability differed in timing and magnitude for the intermittent and perennial streams (Figure 2.3). For the perennial streams and two of the almost-intermittent streams (A2 and A3), the coefficient of variation tended to be below 1.5. Furthermore, only short periods between April and October had relatively high coefficients of variation for discharge (Figure 2.3). The intermittent streams and one almost-intermittent stream (A1), on the other hand, had much higher coefficients of variation that lasted for longer periods in late August and September (Figure 2.3).

Mean annual discharge and the variation in annual discharge, however, were similar for the intermittent and perennial streams (Figure 2.4). Mean annual discharge varied from 73 mm to 245 mm (mean and median 158.5 mm; σ=50.1 mm). Mean annual discharge and mean annual precipitation were correlated (r_s=0.66, p=0.044) but there was no statistically significant relation between mean annual precipitation and runoff ratios (p=0.143) for the 10 stations.

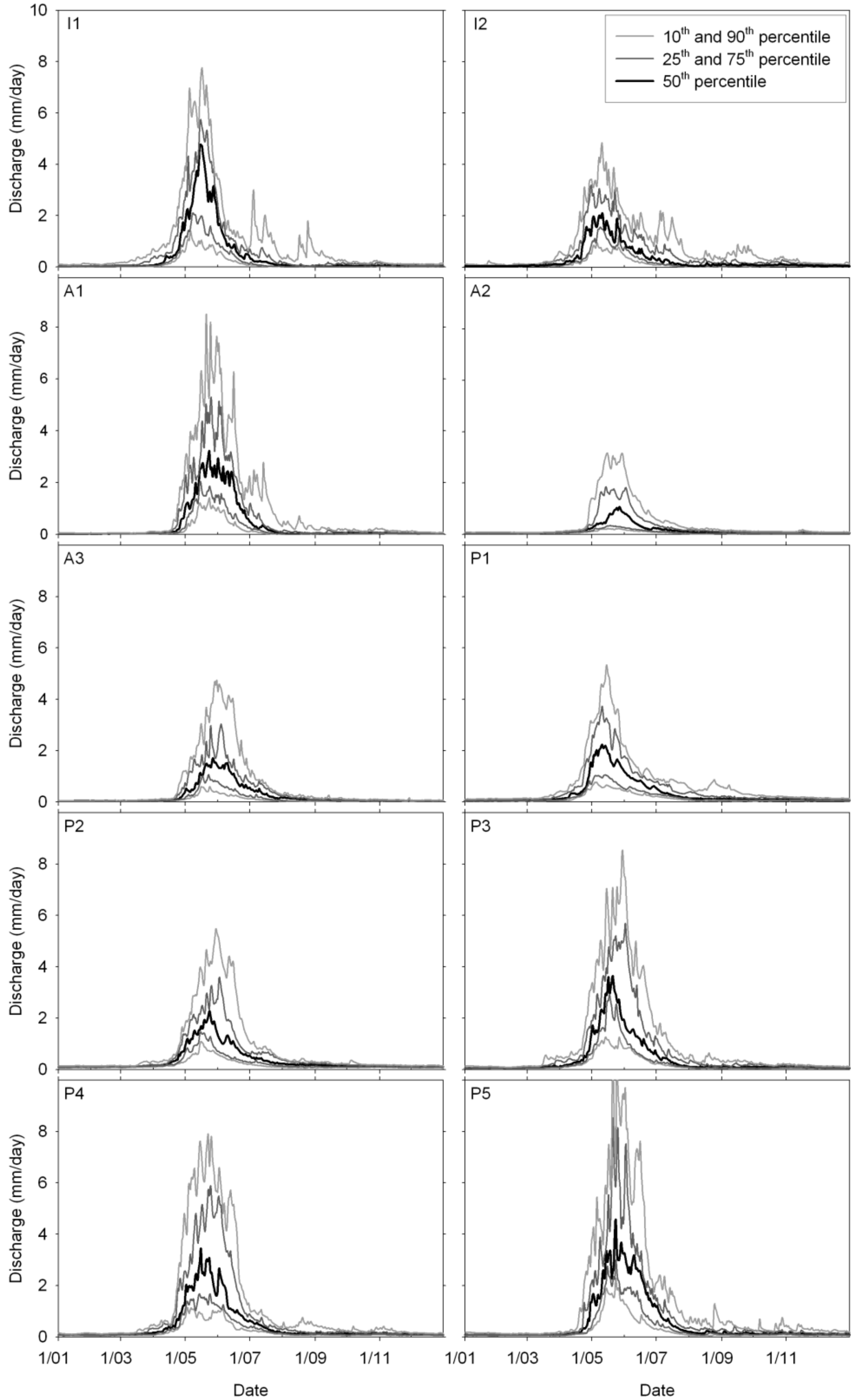

Figure 2.2 **Hydrographs for the 10 streams, showing the 10, 25, 50, 75, and 90th percentile of discharge on each calendar day.**

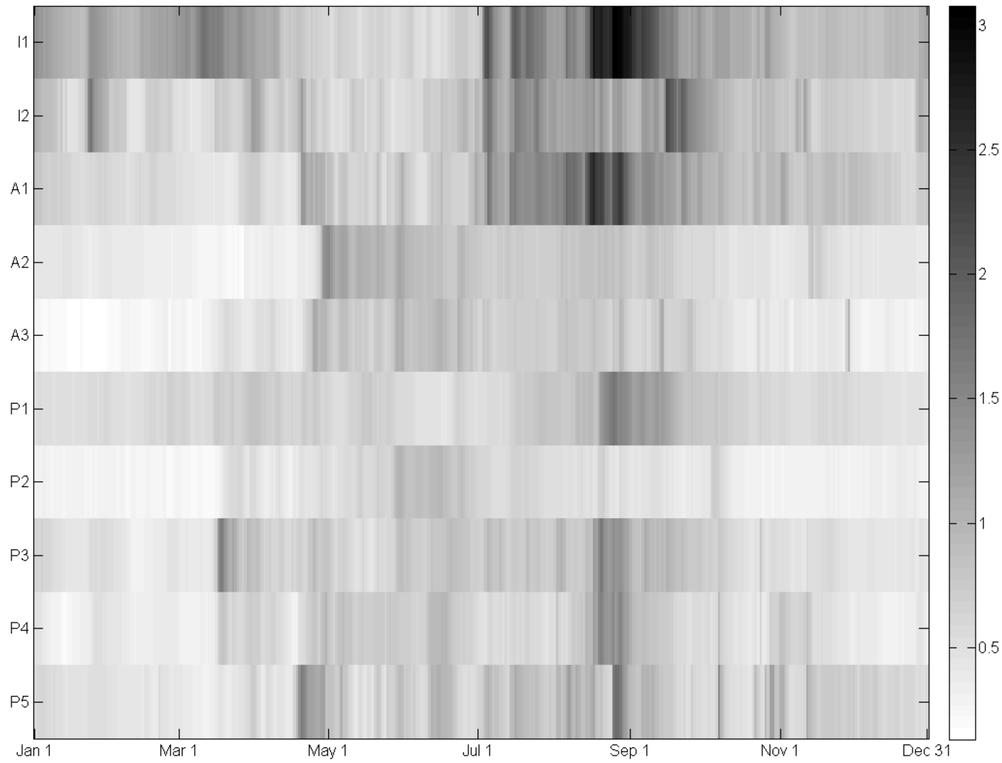

Figure 2.3 The coefficient of variation of discharge on each calendar day, the two top rows (I1 and I2) are the intermittent streams.

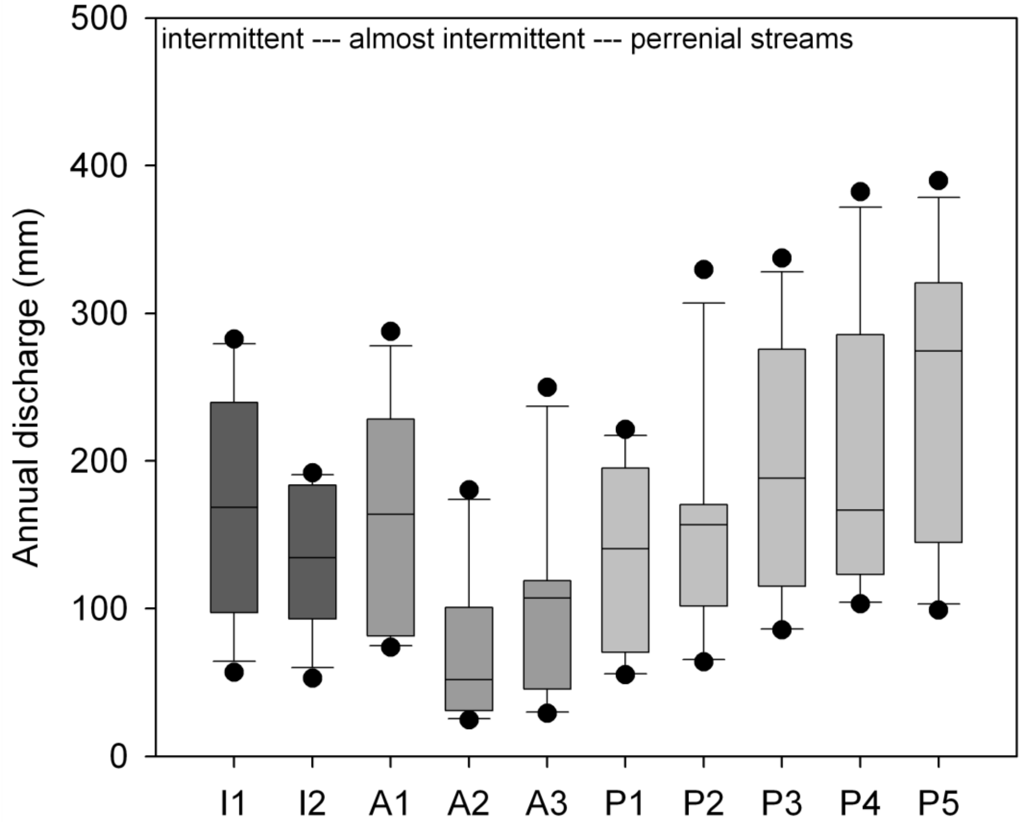

Figure 2.4 **Box plot of the annual discharge of the 10 stations for the 1972-1982 period. The box represents the 25th and 75th percentile, the line the median, the whiskers the 10th and 90th percentile and the dots the outliers.**

2.5.2. Spring freshet

The intermittent streams tended to have a slightly earlier median start of spring freshet compared to the perennial and almost-intermittent streams, although the earliest start date of the freshet did not differ between the intermittent and perennial streams (Figure 2.5). The day of the year of the peak freshet and the median peak freshet, however, were similar for the perennial and intermittent streams (Table 2.2 and Figure 2.2). The median day of the year of the start of the freshet and the peak freshet occurred earlier for watersheds located further north and further west (Table 2.3). Not surprisingly, the median day of the year of the peak of the freshet was also correlated with the average and maximum elevation of the watershed (r_s=0.96 and 0.86, p=10^{-5} and 0.001 respectively; Table 2.3).

Table 2.3 Spearman rank correlation coefficients (r_s) for the relations between the physical characteristics of the watersheds and the streamflow characteristics. Values are only shown for correlations with $p<0.10$. Values in italic represent correlations with $p<0.05$, while values in *bold* represent correlations with $p<0.01$.

	Latitude	Longitude	Mean annual precipitation (mm)	Drainage area (km^2)	Length of streams (km)	Drainage density (km/km^2)	Average slope	Maximum slope	Average elevation (m asl)	Maximum elevation (m asl)	% Forest	%Intrusive	%Metamorphic	%Volcanic	% sedimentary
Mean annual discharge (mm)			*0.66*												
Runoff ratio															
% of Zero Flow days (1972-1982)				*-0.70*	*-0.70*		*-0.68*	-0.62		-0.62	*-0.68*	-0.61	0.56		
Median peak freshet (mm/d)			***0.92***										0.59		
Median summer low flow (mm/d)						*0.71*									
Median peak fall flow (mm/d)											*-0.68*				
Fall rise (mm)											-0.56		*0.65*		
Median June 15 flow (mm/d)				*0.75*	0.62			0.60	***0.83***	***0.88***					
Median August 15 flow (mm/d)															
Median March 15 flow (mm/d)						***0.90***	0.64								0.57
Median DoY of start of freshet		*-0.66*										***0.85***		*-0.76*	
Median DoY of peak freshet	*-0.75*	-0.58							***0.96***	***0.86***		*0.72*		*-0.71*	
Median DoY of summer low flow					0.56		0.61								
FDC_{slope}		*0.74*	***0.78***									*-0.73*	*0.68*		
Master recession constant (k)			*-0.70*			*0.67*							*-0.70*		

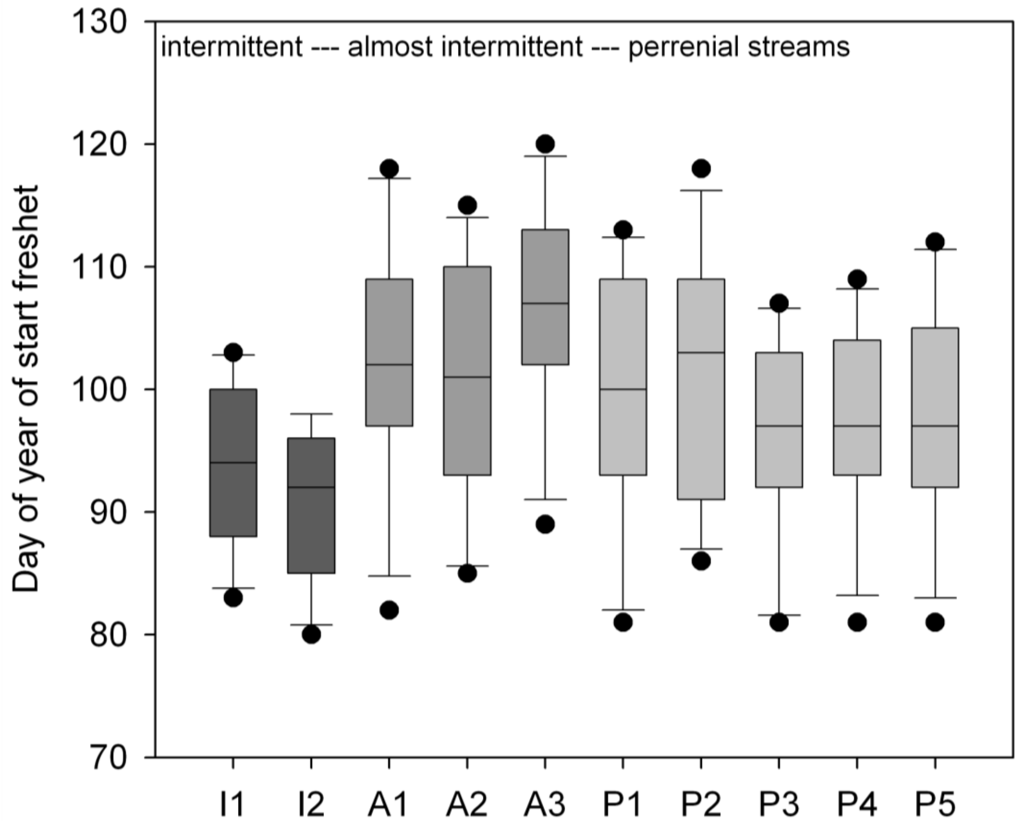

Figure 2.5 Box plot of the day of year of the start of the spring freshet (for explanation of the boxplots see Figure 2.4). DoY 70 represents March 11th and DoY 130 represents May 10th.

2.5.3. Summer low flows

As expected, the intermittent streams and one almost-intermittent stream (A1) had considerably lower summer minimum flows compared to the perennial streams and the other almost-intermittent streams (A2 and A3) (Table 2.2). The intermittent streams were also characterised by an earlier summer minimum streamflow, although there was a large spread in the data (Figure 2.6 and Table 2.2). The median summer minimum flow was only statistically significantly correlated to the drainage density (r_s=0.71, p=0.022) (Table 2.3). The median day of the year of the summer low flow was only correlated to the average slope (r_s=0.61, p=0.061) and stream length (r_s=0.56, p=0.092), with summer minimum flow occurring later for watersheds with steeper slopes (Table 2.3).

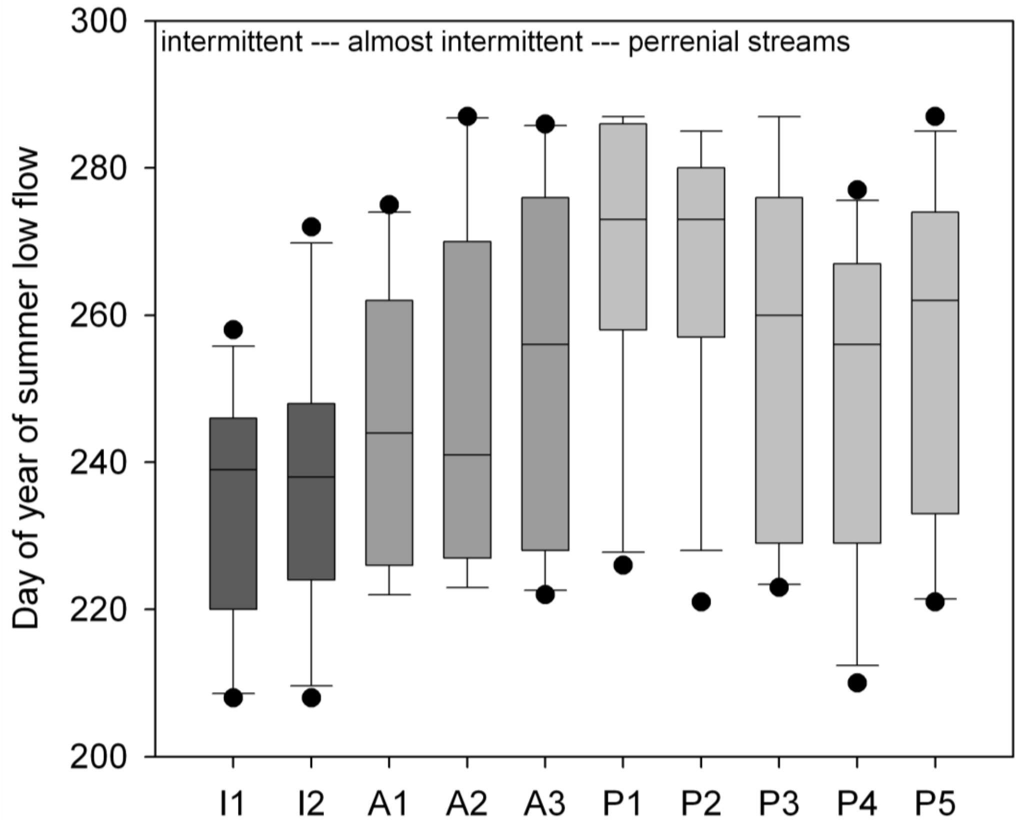

Figure 2.6 Box plot of the day of year of the summer minimum flow (for explanation of the boxplots see Figure 2.4). DoY 200 represents July 19th. DoY 300 represents October 27th.

Interestingly, the intermittent streams and almost-intermittent stream A1 had almost the same median discharge on March 15th and August 15th or a lower median discharge on August 15th than on March 15th, while for the perennial streams and the other almost-intermittent streams the median discharge was significantly higher on August 15th than on March 15th (Table 2.2). Median discharge on August 15th was not correlated to any of the physical watershed characteristics for which we had data. Median discharge on March 15th was positively correlated with drainage density (r_s=0.90, p=0.001), average slope (r_s=0.64, p=0.054) and the fraction covered by sedimentary bedrock (r_s=0.57, p=0.083) (Table 2.3). Three of the four watersheds with the highest median March 15th discharge, were located in the Northern part of the Okanagan Basin and had areas of sedimentary bedrock (Tables 1 and 2). However, watershed P2, located in the western part of the Okanagan Basin, has no sedimentary bedrock and has the highest median March 15th discharge.

A comparison of master recession curves clearly illustrated that the master recession constant was lower for the intermittent streams and almost-intermittent stream A1 than for the other almost-intermittent streams and perennial streams (Table 2.2), indicating a faster decrease in streamflow and storage during the recession period for the intermittent streams and almost-intermittent stream A1 than the other streams. The master recession constant was correlated with drainage density (r_s=0.67, p=0.039) and inversely correlated with the fraction of the watershed that consists of metamorphic bedrock (r_s=-0.70, p=0.025) (Table 2.3). Only the intermittent streams, almost-intermittent stream A1 and P5, which becomes intermittent downstream of the gauging station, are covered in part by metamorphic bedrock (Table 2.1).

2.5.4. Fall peakflows

The intermittent streams had large variations in fall peakflows and tended to have a higher maximum fall discharge than the perennial and almost-intermittent streams (Figure 2.2 and Figure 2.7). The combination of higher fall peakflows and lower summer low flows means that during some years the relative rise in streamflow was much larger for the intermittent streams than for the perennial streams. However, the minimum fall peakflow was also lower for the intermittent streams and almost-intermittent stream A1 than for the other streams (Figure 2.7).

To further understand the significance of fall peakflows and possible regime differences, individual hydrographs were examined. This revealed that high fall peakflows between September 1st and November 1st did not occur every year at any of the streams but occurred more frequently for the intermittent streams than for the perennial streams. Furthermore, watersheds on the eastern side of the Okanagan Basin had higher fall peakflows than those on the drier west side. For example, in 1976 all watersheds with high fall peakflows were located in the east, although P4 and P3 in the northwest also had fairly high fall peakflows, while P2, A3 and A2 in the west had very low fall peakflows or no increases in streamflow in fall at all. Not only the intermittent streams, but also P1 had very high fall peak flows in 1976. This was also true for 1978, when most eastern streams had high fall peakflows and streams on the western side had low fall peakflows.

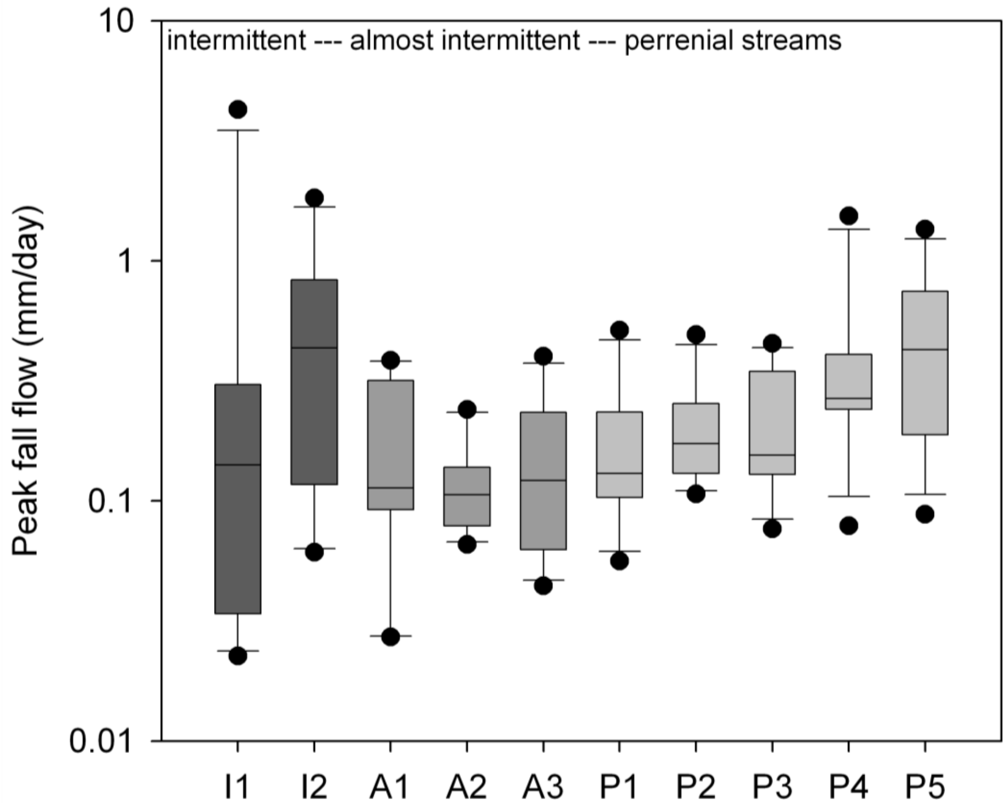

Figure 2.7 Box plot of fall peak flow (for an explanation of the boxplots see Figure 2.4).

The median fall peakflow was inversely related to the area of the watershed covered by forest (r_s=-0.68, p=0.035) (Table 2.3). There was also a correlation between drainage density and median fall peakflow but the two watersheds with the highest median fall peak flows (I2 and P5) did not follow this trend. Therefore, the relation between drainage density and median fall peakflow was not statistically significant (p=0.349).

2.5.5. Flow duration curves

The slopes of the FDCs differed for the intermittent and perennial streams. A qualitative visual examination of the general shape of all ten FDCs indicates that the high discharges follow a similar frequency distribution, with the exception of A2. At the 20th percentile the almost-intermittent streams start to deviate from the general frequency pattern and the FDC slopes steepen through to the 60th percentile, at which point the slopes become similar again to the perennial streams. In contrast, the intermittent streams tend to have a fairly uniform slope, with the exception of the upper and lower reaches of the curve.

The FDCs of the intermittent streams were steeper at low to medium discharge compared to those of the perennial streams (Table 2.2 and Figure 2.8). FDC_{slope} ranged between 4 and 5 for the intermittent streams and almost-intermittent stream A1, whereas for the perennial and remaining almost-intermittent streams it ranged from 1.9 to 3.3 (Table 2.2). These results compare well with the master recession results and also suggest that streamflow declined more rapidly during the summer at the intermittent stations and station A1 than at the other almost-intermittent stations and the perennial stations. FDC_{slope} was correlated with the percentage of time with zero flow (r_s=0.67, p=0.036). It was also correlated with longitude (r_s=0.74, p=0.014), the fraction of metamorphic bedrock (r_s=0.68, p=0.030), and inversely correlated to the fraction of intrusive bedrock (r_s=-0.73, p=0.017) (Table 2.3). Longitude and the fraction of the area covered with metamorphic bedrock (r_s=0.57, p=0.083) and intrusive bedrock (r_s=-0.69, p=0.027) are also correlated.

2.5.6. Cluster analysis

Cluster analysis showed that based on the physical watershed characteristics, the ten watersheds can be divided into five groups (Figure 2.9a-b). The intermittent streams (I1 and I2) form one group and are thus distinct from the other watersheds. P5, which becomes intermittent below the station, forms itself a group. These two watershed groups are characterised by a low average slope and the smallest fraction of the watershed covered by forest. P1, is the only watershed that consists predominantly of sedimentary bedrock (Table 2.1), also is distinct from the other groups. The almost-intermittent streams A2 and A3 form another group together with perennial stream P2. These watersheds are located on the west side of the Okanagan Basin and are for a large part composed of intrusive rock (Figure 2.9, Table 2.1). The final group consists of almost-intermittent stream A1 and perennial streams P1, P3, and P4. These watersheds have the lowest minimum elevations.

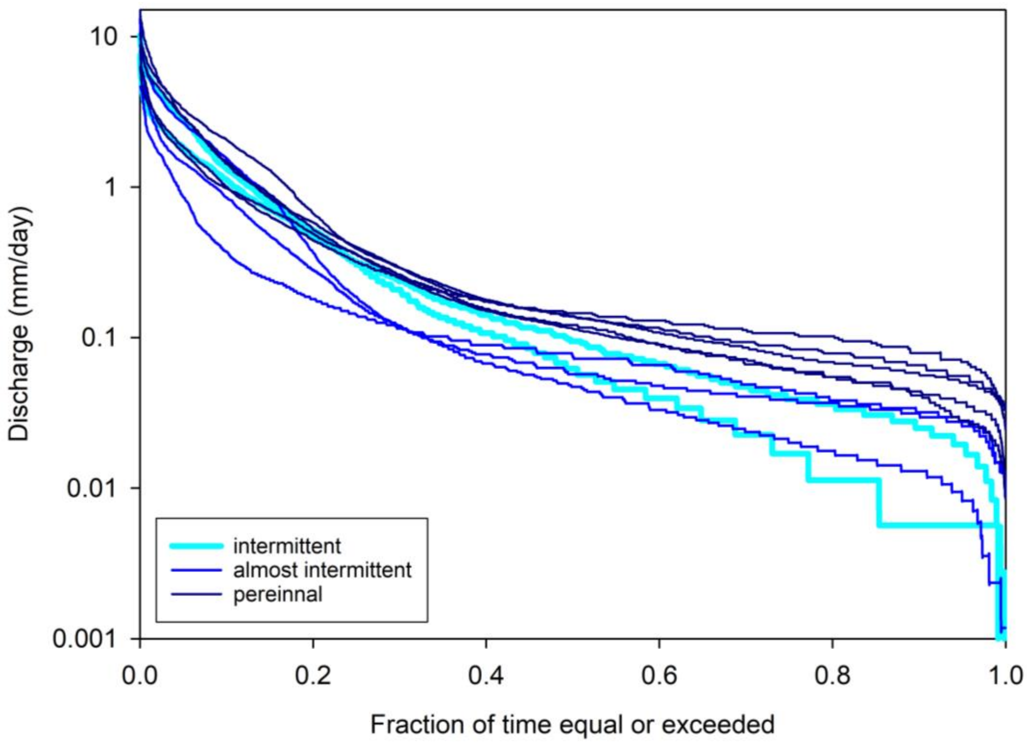

Figure 2.8 Flow duration curves for the ten streams.

The cluster analysis based on the main hydrograph characteristics (Table 2.2) showed that there are three types of hydrographs: those from the intermittent streams plus A1, from the other almost-intermittent streams A2 and A3, and from the perennial streams (Figure 2.9c). Thus, even though the behavior of almost-intermittent streams A2 and A3 appeared similar to those if the perennial streams, the cluster analysis results suggest that they are distinct from the perennial streams, and distinct from the intermittent streams and A1.

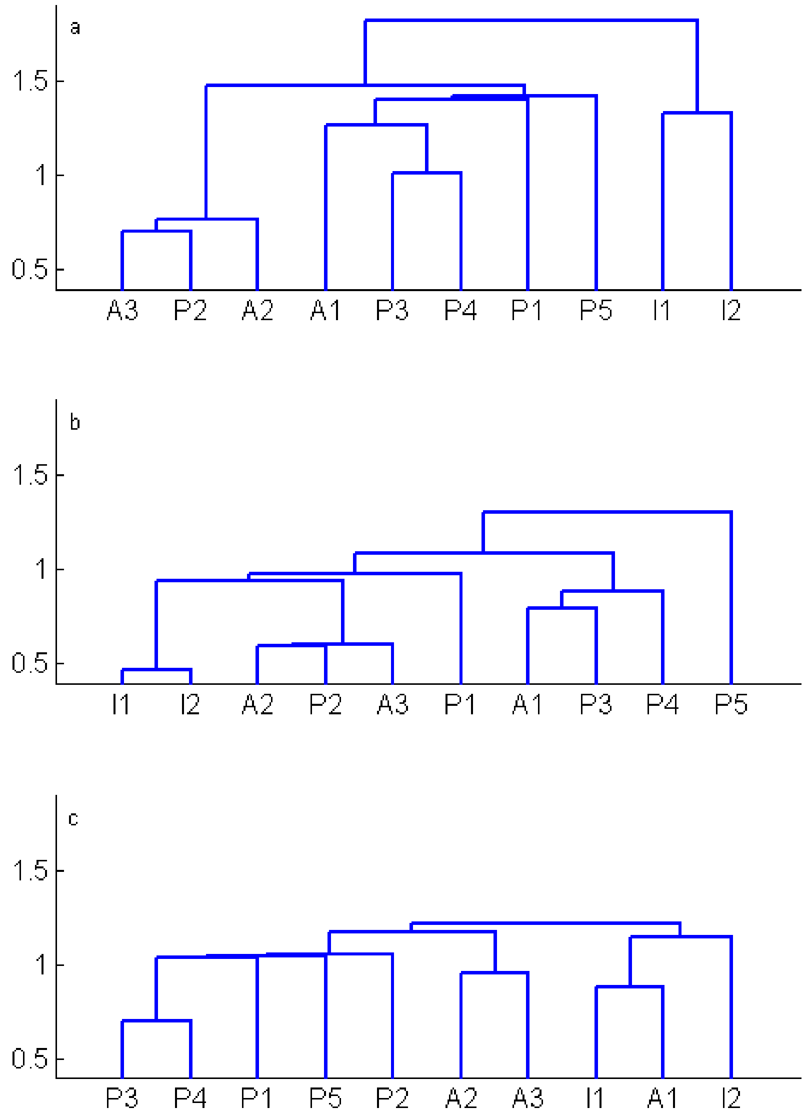

Figure 2.9 **Results of the cluster analysis based on all physical characteristics (latitude, longitude, drainage area, drainage density, average and maximum slope, average, minimum, and maximum elevation, fraction covered by forest, fraction of intrusive bedrock, metamorphic bedrock, volcanic bedrock, and sedimentary bedrock) (a), based on only the topographic and location characteristics (latitude, longitude, drainage area and density, average and maximum slope, average, maximum, and minimum elevation (b), and the main hydrograph characteristics (mean annual flow, mean runoff ratio, median peak freshet, median summer low flow, median fall peak flow, median day of year of start of freshet, peak freshet, and summer low flow, median flow on March 15 and August 15, FDC_{slope}, and the master recession constant) (c).**

2.6. Discussion

The five streamflow regime components of Poff and Ward (1989) to characterise watersheds were used to assess the degree of similarity and dissimilarly between intermittent, almost-intermittent, and perennial streams in the Okanagan Basin. All ten studied watersheds are characterized by nival (snowmelt dominated) runoff regimes (Figure 2.2). The streamflow hydrographs (Figure 2.2) are similar in shape: a steep rising limb to peak discharge in May, followed by a long recession. There are no obvious visual differences in the overall shape of the median hydrographs between the intermittent, almost-intermittent and perennial streams. This similarity in the hydrograph shape is not unexpected as all watersheds are affected by the same climatic regime. It is thus the subtleties of the hydrographs that indicate that there may be differences between intermittent, almost-intermittent and perennial streamflow regimes in the Okanagan Basin. Almost-intermittent stream A1 exhibited many streamflow characteristics that are similar to the intermittent streams but the cluster analysis results suggest that its physical characteristics are distinct from those of the two intermittent watersheds. The other two almost-intermittent streams had flow regimes that resembled the perennial streams, although cluster analysis puts them separate.

Magnitude

Annual discharge

There were no clear differences in either the mean or median annual discharge between the intermittent, almost-intermittent, and perennial streams (Figure 2.4 and Table 2.2). There were no differences in peak freshet between the intermittent, almost-intermittent, and perennial watersheds either, most likely because all watersheds experienced similar snow accumulation and snowmelt regimes.

Summer low flows

Not surprisingly, median summer low flows were significantly lower for the intermittent streams and almost-intermittent stream A1 than for the other almost-intermittent streams and the perennial streams. The average of the median summer minimum discharge for the intermittent streams and stream A1 was only 20% of the average of the median minimum summer discharge of all ten stations. Summer low flows are often determined by groundwater discharge into the stream and geomorphology (Smakhtin, 2001; Tague and Grant, 2004; Price et al., 2011). Although recession constants were high for all watersheds, they were lowest for the intermittent streams and the FDCs were steepest for the intermittent streams (Table 2.2), indicating that the intermittent streams likely have less groundwater inflow and/or had groundwater reservoirs that drained more quickly. However, field observations in England by Anderson et al. (1980) showed that at very low flows, discharge can be maintained by throughflow in the vadose zone, especially in watersheds with impermeable bedrock. For these situations they reported recession constants greater than 0.9 and yet found no evidence of actual groundwater contribution to streamflow.

The watersheds of the intermittent streams have the lowest slope and drainage density and may therefore have fewer points of interaction with regional groundwater flow paths. The 2-dimensional modeling results of Neilson-Welch et al. (2011) show that bedrock groundwater discharge was less for the hillslope with the lower slope, except for the very low applied recharge case. This was attributed to the longer seepage face and shorter upslope recharge zone for the model with the lower slope. Other studies have shown that watersheds with low drainage densities are characterized by more bedrock groundwater flow. Onda et al. (2006), for example, showed for watersheds in Japan that bedrock groundwater flow was dominant in un-weathered hard shale and argillite watersheds, whereas subsurface stormflow through the soil mantle was the dominant runoff generation mechanism in granite watersheds that were not as deeply dissected. Tague and Grant (2004) showed that summer streamflow was higher and recessions were slower for watersheds with a higher percentage of High Cascade geology, dominated by low gradient basaltic and andesitic lava flows, cinders, pumice, and volcanic ash, that are characterized by low drainage densities (1–2 km/km2). Unfortunately, we have no information on the characteristics of the bedrock or the regional groundwater flow paths in the Okanagan Basin and thus cannot determine the effects of slope and drainage density on groundwater contributions to streamflow.

Winter low flows (e.g. March 15th flow) and summer low flows (e.g. August 15^{th} flow) were similar for the intermittent streams and almost-intermittent stream A1, whereas low flows were significantly larger in summer than in winter for the perennial streams and the other two almost-intermittent streams. Like the results from the flow duration curves and the master recession curves, this also suggests that groundwater storage is depleted faster for the intermittent streams and almost-intermittent stream A1 than for the other streams.

March 15th flows were highest for watersheds with sedimentary bedrock that are located in the Northern part of the Okanagan Basin, have high drainage densities and low minimum elevations. Winter rainfall and rain-on-snow melt events at lower elevations may possibly contribute to the larger winter flows for these watersheds. However, winter temperatures were not higher than the average of the 10 watersheds, suggesting that differences in precipitation (rain vs snow) are small. It is thus likely that the high March 15th flows are at least in part related to differences in groundwater flow pathways, which is in line with the results of Onda *et al.* (2006), who showed that bedrock groundwater flow was more important in watersheds underlain by shale than in watersheds underlain by granite. However, this is contrary to the results of Shimizu (1980, referenced in Tani *et al.*, 2012) and Katsuyama (2008), who showed that watersheds underlain by sedimentary bedrock had a more flashy runoff response and less sustained baseflow than watersheds underlain by igneous bedrock.

Fall peakflows

Fall peakflows were most variable for the intermittent streams. They were low for the intermittent streams in some years, but in years with high fall precipitation, they were highest for the intermittent streams, resulting in a much larger variability in fall peakflow for the intermittent streams than the perennial streams (Figure 2.5). The effects of localized large storms are larger in small watersheds than in large watersheds. However, drainage area does not explain the differences between median fall peak flow for the watersheds studied here. Watershed I2 is similar in size to A2 and P2, and the drainage area of A1 is similar to that of the other perennial watersheds. There was no significant correlation between drainage area and median fall peakflow ($p=0.973$) either and high fall peakflows occurred in several watersheds, suggesting that it was not likely due to localized precipitation events. Furthermore, Toews et al. (2009) showed that local precipitation in the Okanagan mainly occurs between April and September and that recharge from these localized precipitation events is insignificant.

The higher fall peakflows during the years with high fall rainfall could be caused by a lower soil moisture and groundwater storage capacity for the watersheds with intermittent streams or faster and more direct flow pathways, which would also cause them to have lower summer low flows and an earlier start of the freshet. Median fall peakflow and median day of the start of the freshet (r_s=-0.62, p=0.056) were correlated. However, we currently do not have any information on the storage and hydraulic properties of the different watersheds and thus cannot confirm if the watersheds of the intermittent streams have a lower storage capacity and/or higher hydraulic conductivity soils with shorter and faster flow pathways. Watersheds with higher drainage densities and shallower flow paths generally have a flashier response than watersheds with a lower drainage density (Tague and Grant, 2004; Onda *et al*., 2006). However, in this study the watersheds of the intermittent streams that had the highest fall peakflows, have the lowest slopes and lowest drainage densities.

The median fall peakflows was highest for the watersheds with the lowest forest cover (Table 2.3). Reduced forest cover leads to lower interception and transpiration losses. As a result, soils are wetter and respond more quickly to fall recharge events. Paired watershed experiments have shown that streamflow increases after logging are frequently largest in the fall (Moore and Wondzell, 2005) but because the watersheds with relatively low forest cover still had more than 50% forest cover and in years without high fall precipitation had the lowest maximum fall discharge, it is not likely that this is the main cause for the differences in fall peakflows.

2.6.2. Frequency and duration

The streamflow frequency regimes for the ten watersheds indicate there are differences and similarities between intermittent and perennial streams. This was expected as the shape of the FDC is partially defined by climate and watershed characteristics (Searcy, 1959; Vogel and Fennessey, 1995), which are somewhat similar throughout the Okanagan Basin. The frequency and duration of zero flow for the intermittent streams was minimal (Table 2.1). The intermittent periods occurred during the late summer (August-September) and persisted for a maximum of ten consecutive days. Zero flow was recorded at I1 in two of the eleven years (1973 and 1978) and at I2 during only one year (1979), although discharge was frequently below 0.01 mm/day for a week or more during August or September for I1, I2, and A1. Discharge was rarely, if ever, below 0.01 mm/day for the other almost-intermittent and perennial streams. Periods of zero flow are more frequent for the truly intermittent streams, which are not gauged continuously and are therefore not part of this analysis.

The FDCs were steeper at low and medium discharge for the intermittent streams compared to the perennial streams (Table 2.2 and Figure 2.8). Had there been data for truly intermittent streams with longer periods of zero flow, the shape of the lower end of the FDCs would have been even steeper and cut off earlier. The conceptual framework for reconstructing FDCs suggests evapotranspiration (ET) and soil moisture storage capacity control the slope of the FDC at low discharges, while high ET from the saturated surface and/or low soil storage capacity steepen the FDC (Yokoo *et al*., 2011). This suggests that evapotranspiration rates are higher in the watersheds of the intermittent streams and/or that these watersheds have a lower soil storage capacity than the watersheds with the perennial streams.

The slope between the 33rd and 66th percentiles of the FDC ranged between 4 and 5 for the intermittent streams and almost-intermittent stream A1, whereas for the perennial and remaining almost-intermittent streams the slope ranged from 1.9 to 3.3 (Table 2.2). The slope between the 33rd and 66th percentiles of the FDC is controlled predominantly by groundwater or subsurface flow. Steeper FDC slopes suggest less groundwater discharge to the stream and more permeable soils (Searcy, 1959; Vogel and Fennessey, 1995; Yokoo and Sivapalan, 2011). These results indicate a larger groundwater influence in perennial streams compared to the intermittent streams and are consistent with the master recession results. Field research is required to quantify and validate these findings.

Discharge above the 10th percentile were all within a factor of two, with the exception of A2 which had a significantly lower discharge at the upper end of the FDC. The dominant control on the upper portion of a FDC is precipitation (Yokoo and Sivapalan, 2011). The similarity in shape of the upper portions of the FDCs is thus likely explained by the relatively similar input of melt water during freshet. There is no obvious explanation for the low spring freshet in watershed A2.

2.6.3. Timing

The freshet tended to start slightly earlier for the intermittent streams, although the earliest day of the year of the start of the freshet in the 11 years of record was similar for the intermittent, almost-intermittent and perennial streams (Figure 2.5). An earlier start of the freshet can often be explained by lower elevations and or more southerly aspects causing an earlier onset of snowmelt, but the mean elevation of the watersheds with the intermittent streams was neither the highest nor the lowest. Watershed I1 has the second highest minimum elevation, which would normally be reason for a later start of peak freshet; neither watershed I1 nor I2 had the lowest maximum elevation. In fact, there was no correlation between the median day of year of the start of spring freshet and elevation ($p=0.723$, 0.208, and 0.287 for minimum, average, and maximum elevation respectively). The median day of peak freshet was highly correlated with the average and maximum elevation of the watershed (Table 2.3).

The correlation between the median day of the year of the start of the freshet and the area of intrusive and volcanic bedrock (Table 2.3) suggests that perhaps geomorphological aspects of these watersheds or soil properties influence the start of the spring freshet. However, we currently do not have enough information on these watersheds to investigate this in more detail. As the median day of the year of the start of the freshet occurred only slightly earlier for the intermittent streams (Table 2.2), it could also be related to the hydraulic conductivity of the soils, their storage capacity, and flowpath-lengths through the watershed. Perhaps melt is initiated at all watersheds around the same time but shallow soils or an impermeable layer beneath the surface causes water to flow to the stream faster in the watersheds with intermittent streamflow. Shallower flow paths or more conductive soils could also explain the larger increases in streams in fall for the intermittent streams.

2.6.4. Rate of change

The intermittent streams had much higher fall peakflows during years with high fall precipitation and thus a much higher rate of change in streamflow in fall during those years than the other streams. However, in years without large amounts of precipitation in the fall, maximum fall discharge was very low and the rate of change in fall was thus very low. This results in a higher variability (Figure 2.3) and a much lower predictability in streamflow in fall for the intermittent streams than the perennial streams. This higher rate of change and lower predictability in fall peakflow could have large ecological consequences (Poff *et al*., 1997).

Leith and Whitfield (1998) examined the streamflow records of six streams in south-central BC for hydrological changes that may be caused by climate change. They found that the temporal pattern of streamflow has changed and that the changes were consistent with those expected from climate change. The studied streams exhibited an earlier freshet, lower late summer flows, and earlier and higher fall flows during the 1984-1995 period than during the 1971-1983 period. These results are attributed to higher temperatures and thus earlier melt in spring causing a longer dry season, and precipitation in fall being rain instead of snow. These changes in streamflow characteristics correspond with the subtle differences between the intermittent and perennial streams in this study. The intermittent streams and almost-intermittent stream A1 had a slightly earlier median day of the year of the start of the freshet, a lower summer minimum streamflow, and higher fall peak flows during some years than the perennial streams. This thus suggests that these streams may be most vulnerable to changes in climate. Because Leith and Whitfield (1998) showed that these changes occurred after 1983 and for this study only data until 1982 was used, it is important to also study current differences between intermittent and perennial streams and whether some of the former perennial streams now have streamflow regimes that resemble those of the intermittent streams or if the differences between the intermittent and perennial streamflow regimes have become more pronounced.

The recession constants were lowest for the intermittent streams and the FDCs were steepest for the intermittent streams (Table 2.2), indicating that in addition to the larger rise in streamflow in fall in some years, streamflow in summer also decreased faster for the intermittent streams. This is likely because of less groundwater inflow and/or groundwater reservoirs that drained more quickly, resulting in similar or lower streamflow on August 15th than March 15th for the intermittent streams, while streamflow was significantly higher on August 15th than on March 15th for the perennial streams.

2.7. Conclusion

This study focused on streamflow regimes of intermittent and perennial streams in the Okanagan Basin. The majority of the analyses showed only subtle differences between intermittent and perennial streams. Most previous studies that have compared runoff regimes have been conducted over areas that are much larger than the Okanagan Basin, where climate is more variable and has a larger influence on the differences in the flow regimes. The differences in the flow regimes and physical characteristics in this study may also only be subtle because our analyses included only two intermittent streams. Differences between intermittent and perennial streams may become clearer (or disappear) if more streams were studied. It is likely that the differences in the flow regimes would be larger if truly intermittent streams in the Okanagan Basin that run dry every single year were included. However, there are currently no year-round data for more than 2 years for these streams. We thus highly recommend continuous monitoring of intermittent streams in order to characterize their behaviour and understand how their flow regimes are different from perennial streams.

Due to the small number of intermittent streams, no strong conclusions about the differences in flow regimes between intermittent and perennial streams can be made but the results suggest that some of their characteristics, in addition to the absolute minimum flow, may be different. The intermittent streams tended to exhibit an earlier start of the spring freshet and higher fall peak flows during years with high precipitation in fall. As a result, they exhibited higher variability in streamflow in late summer and fall than the perennial streams. The intermittent streams also had the steepest flow duration curves and the lowest master recession constants. The difference between the median flow on August 15 and March 15 was also much smaller for the intermittent streams than the perennial streams. These subtle differences in the streamflow regime characteristics, in addition to the lower minimum summer flows, may have important implications for stream ecology and point to different runoff generation mechanisms and residence times for water in the intermittent and perennial watersheds. The slightly earlier start of the freshet, faster decline in streamflow in summer, and faster and larger response to fall precipitation events suggests that perhaps flow pathways are shorter for the intermittent watersheds. The steeper flow duration curves and lower master recession constants suggest that there is less groundwater contribution to the intermittent streams and storage is depleted faster and/or is smaller. These hypotheses need to be confirmed by field data and/or modeling studies.

2.8. Acknowledgements

We would like to thank Judy Kwan (Environment Canada) for sharing the watershed shape files with us, John Ng (Simon Fraser University) for help with Figure 2.1 and obtaining the watershed physical characteristics. I never would have started this thesis without Paul Whitfield, who piqued my interest in intermittent streams. This work was supported by an NSERC Discovery Grant.

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Chapter 3. **Spatial distribution of streamflow along an intermittent stream in the southern Okanagan Basin**

3.1. Abstract

Intermittent streams are understudied but could provide a significant amount of water to downstream perennial water bodies, particularly in arid and semi-arid basins, such as the Okanagan in British Columbia Canada. An improved understanding of the spatial variability in streamflow and the connections between surface-water and groundwater along these stream is vital for effective management of water resources, habitat management and stream ecology in these unique systems. Furthermore understanding these interactions is a pre-requisite to predicting changes in water availability.

This study investigates temporal and spatial variability in streamflow gains and losses along Long Joe Creek. Differential streamflow gauging was used to map streamflow gains and losses and to study the connection and disconnection of flowing stream segments. Temperature measurements provided additional information about the presence or absence of water and the source of the water. The results suggest that the valley fog line, which is located above a change in geology, may have a significant effect on the variability in streamflow. A recent groundwater isotope study suggested there is no groundwater recharge from headwater streams such as Long Joe but the results of this hydrometric study suggest otherwise.

Keywords: Okanagan Basin, intermittent stream, spatial distribution, gaining and losing stream sections

3.2. Introduction

3.2.1. Importance of temporary streams

Temporary streams include ephemeral and intermittent streams (Buttle et al. 2012) and are characterized by a lack of water and flow during at least a part of the year. At least a third of all rivers and streams on the earth's surface are temporary (Buttle et al. 2012; Poff 1992). They range from small headwater streams in humid climates to larger rivers in arid climates. It has been estimated that 59% of the total stream length In the United States (excluding Alaska) consists of temporary streams; in arid or semi-arid environments over 81% are intermittent or ephemeral in nature (Levick et al, 2008). Unfortunately, there are no comparable data for the presence of temporary streams in Canada but temporary streams in Canada include headwater streams, streams in drier areas, like the Okanagan in British Columbia, and streams that stop to flow during extended winter dry periods (Buttle et al. 2012).

Despite their abundance, temporary streams remain relatively understudied as the majority of research to date has focused on perennial rivers. Temporary streams can be an important source of water. Transmission losses from intermittent and ephemeral streams are an important groundwater recharge mechanism in arid and semi-arid regions (Rushton 1997). Understanding the losses from streamflow to groundwater is crucial for the management of water resources these environments (de Vries and Simmers 2002). At the same time groundwater inputs (gains) provides the source water for wet or flowing areas of intermittent channels and is important for maintaining aquatic communities.

Currently, there is limited knowledge on when and where intermittent and ephemeral streams flow because intermittent and ephemeral streamflow is very difficult to quantify and the majority of the gauging stations are located in perennial streams (Chapter 2). Most stream gauges are installed for water management purposes (Kirchner 2006) and where intermittent streams are gauged, they are often only gauged for part of the year and the occurrence of zero flow is not recorded (Peters et al. 2012).

The inherent erratic nature of intermittent flow does not lend well to conventional monitoring techniques. Conventional gauging methods require a stable control section to define the relationship between stage and discharge. Temporary stream channels tend to have a very unstable geomorphology and therefore an unstable control section. Furthermore, the spatial variability in the amount and occurrence of flow in temporary streams is high, so that where you put the stream gauge significantly affects your measurements on how much flow and when flow occurs. Annual streamflow in intermittent streams may decrease with basin area, unlike basins in temperate areas with continuous flow where streamflow generally increases with drainage area (cf Walker et al. 1995). Reach scale controls on the occurrence and amount of streamflow include the amount and distribution of precipitation, hydrogeology, and streambed properties (i.e. the characteristics of the channel and alluvium).

The spatial variability of streamflow is very pronounced for intermittent streams, where sections of the stream go dry and other sections continue to flow, and has important implications for aquatic biota (fish). Whether there is flow or no flow and what sections are connected has a large effect on the survival and distribution of stream biota along a stream network. Limnologists and ecologists have researched temporary stream channels since the 1980s and have shown that temporary streams are unique habitats (e.g.: Boulton & Suter, 1986; Uys & Keeffe, 1997; and Williams & Bonell, 1988). An understanding of the occurrence of flow and the factors that determine the spatial distribution of flow in temporary streams is important for studying the linkages between the hydrology and ecology in these very unique systems, as well as for the effective management of water resources in the future.

Even though there have been a limited number of studies on how the flowing stream network expands and connects seasonally or during events, most studies have shown that the flowing stream network expands and becomes more connected during wetter conditions. There are three main modes how intermittent streams contract (or expand) (Bhamjee et al., 2016; Goulsbra et al., 2014; Goulsbra et al., 2009; Peirce & Lindsay, 2015; Hunter et al. 2005):

- Top-down: because of the smaller contributing areas (and sometimes shallower soils) the upper areas may dry first. Decreasing water tables also lead to shrinking of the flowing stream network from the top down
- Bottom-up: due to higher evapotranspiration and lower precipitation in the valley bottoms compared to the higher elevation areas, combined with

infiltration of the streamwater in permeable streambed material, or due to breaks in slopes and porous allouvial fans.

- Coalescence of drying or flowing sections: Where there are groundwater inputs and losing sections along the stream network or parts where the flow disappears below the bed and reappears further down. Distinct groundwater inputs can be related to preferential flow pathways, bedding planes or fractures. Interbasin groundwater flow (Welch et al., 2012) can lead to distinct groundwater inputs and more prolonged streamflow at distinct locations as well.

How streamflow varies along stream networks provides useful information on the balance of inflows and outflows (Godsey and Kirchner 2014), and therefore hydrologic flow pathways, such as, inflows from deep groundwater (Huff, et al., 1982), hillslopes (Anderson and Burt 1978) and hyporheic inflows (Schmadel et al. 2017). Beven (2006) stated "*We need more studies of the incremental discharges into stream channels, so that we are encouraged to explore the reasons for the heterogeneities in inputs found*" (p. 396).

3.2.2. Determination of the occurrence of flow in temporary streams

So far, determination of the occurrence of flow in temporary streams has relied mainly on direct observations (i.e. mapping) (Godsey & Kirchner, 2014; Jensen et al., 2017; Shaw et al., 2017; Whiting & Godsey, 2016; Zimmer & Mcglynn, 2017), temperature measurements (e.g. Constantz et al. 2001), conductivity measurements (Bhamjee & Lindsay, 2011; Chapin et al., 2014; Goulsbra et al., 2014) and differential gauging (e.g. (Payn et al. 2009). While direct observations (and mapping) provide the most reliable information on the occurrence of flow, this can only be done for a limited number of streams. It is very time consuming to survey an entire stream to determine the seasonal variation in streamflow.

Temperature can be used to determine timing and duration of temporary streamflow (e.g. Constantz and Thomas 1996; Constantz et al. 2002; Blasch et al. 2004). Temperature based monitoring of the occurrence of flow is based on the assumption that the stream temperature follows the air temperature when the stream is dry and diurnal variations are damped when there is water. Constantz et al. (2002) used temperature profiles to quantify streambed percolation rates and compared them to surface channel losses to determine their accuracy. They found that percolation accounted for 30%-50% of the total channel losses, while evapotranspiration accounted of approximately 10% and the remaining losses were due to the non-vertical component of shallow groundwater flow. Temperature can also be used to determine the variations in infiltration beneath ephemeral streams (Ronan et al. 1998). Percolation rates have been estimated using a numerical heat and water transport model (VS2DH) (Ronan et al. 1998; Constantz et al. 2001; Dowman et al. 2003, and Skinner 2006). These studies all focused on temperature data from select locations and then extrapolated the results across the channel length. Longitudinal temperature profiles along the steambed can provide insight into source water and water exchange zones along the stream. Reaches with cool-water have been observed in areas with upwelling hyporheic water (e.g., Bilby 1984) and/or groundwater (e.g. Story et al. 2003). A disadvantage of temperature profiles is their inability to capture the multidimensional spatially variable nature of streamflow losses beneath losing streams. Other drawbacks are that when these measurements are used in small headwater channels, they do not allow differentiation between pooling water or flowing water, water can flow around the sensor, or that the probe can be buried in sediment. Additionally, temperature profiles are difficult or impossible to install in shallow soils or bedrock. The advantages of temperature profiles are their ability to obtain continuous data on the patterns of intermittent flow and that they are generally considered robust.

Electrical conductivity (EC) sensors can also be used to determine the spatial and temporal distribution of streamflow. EC sensors or state loggers assume that the conductivity increases when there is water (e.g., (Bhamjee & Lindsay, 2011; Chapin et al., 2014; Goulsbra et al., 2014; Goulsbra et al., 2009) but one doesn't know if it is pooling or flowing water (pools of standing water or pools during rainfall events), sensors can get buried in wet sediment and water can flow around the sensor.

Differential gauging, i.e., measuring streamflow at two locations and determining the difference, has been used successfully to determine gains and losses in stream channels (Harte & Kiah, 2009; Langhoff et al., 2006; Mccallum et al., 2014; McCallum et al., 2012; Opsahl et al., 2007; Ruehl et al., 2006). This approach also only provides the net gains or losses over the channel reach. A drawback to differential gauging is that it is only possible at a few locations due to the difficulties in gauging intermittent streams mentioned before (high sediment load, problems with representativeness of the sites where you measure, the accuracy of both measurements to determine the (small) difference, etc.). Covino and McGlynn (2007) used water level and conductivity transects across an entire catchment. This approach reduces errors in scaling reach measurements but does not measure temporal variations in streamflow.

3.2.3. Objectives of this study

In order to further expand our knowledge on the patterns of flow along intermittent streams in headwater catchments, we studied the occurrence and distribution of streamflow along an intermittent stream in the Okanagan, BC. Intermittent streams are common in the Okanagan but are so far largely unstudied. Particularly, we focus on how streamflow changes along the stream in order to determine how intermittent streams in the Okanagan dry up after the freshet and how connectivity is re-established in fall. The specific objectives of this study were to determine:

1) The occurrence and distribution of streamflow in the Long Joe catchment and relate this to topographic or catchment characteristics (accumulated area, slope, bankfull discharge, bed material, elevation, and presence of flow in the winter)

2) The usefulness of surface stream temperature measurements for determining when and where intermittent headwater streams go dry by comparing them with observations of intermittent streamflow along the stream network

3.3. Study site

3.3.1. Regional setting: the Okanagan Basin

The study catchment, Long Joe Creek, is located in the southern portion of the Okanagan Basin, east of Osoyoos, British Columbia. The Okanagan Basin is approximately 180 kilometers in length, with headwaters separating the drainage basins of the Columbia and Fraser Rivers, and extending south to the Columbia Plateau, in Washington State. The lake and river system that dominate the valley floor are primarily fed by small tributary streams located in narrow incised steep valleys. The main channel and a few perennial side channels are monitored year-round by government agencies. The streamflow regime of near-intermittent streams in the Okanagan is slightly different from the perennial streams (Chapter 2) but long-term records for truly intermittent streams are not available.

Being located in the rain shadow of the Coast and Cascade Mountains, the climate of the Okanagan is arid to semiarid. The Okanagan Valley is the driest region in southern Canada. Evapotranspiration is the main component of the water balance: approximately 85% of precipitation in the Okanagan is lost through evapotranspiration and lake evaporation (Cohen et al., 2001). The Okanagan has undergone rapid human population growth and an intensification of agriculture (orchards, vineyards, cropland, and pastures), increasing the pressure on water resources; it has the lowest per person water availability in Canada (OBWB, 2017 (http://www.obwb.ca/wsd/)). The Okanagan's economic dependence on farming (irrigation), forestry and water-based tourism makes it very vulnerable to shortages in water supply. Agriculture accounts for 55% of water use in the Okanagan (Summit Environmental Consultants Inc. 2010) and agricultural demand is expected to rise 24%-38% by the 2050s due to a lengthening of the growing season and increased evapotranspiration (Neilsen et al., 2006). Droughts are not uncommon in the region. During the 2003 and 2004 drought there were water shortages and major fires.

Climate change will further reduce water availability in the Okanagan and increase demands on water (Langsdale et al. 2009). Harma (2010) used an integrated water management model and found that by 2050 in normal and dry years streamflow will be less than demand and there will be a reduced ability to meet instream and downstream demands. There will also be an increasing number of extreme events, including droughts (IPCC, 2007). This will likely increase the number of intermittent streams and the duration of flow intermittency. The IPCC (1997) indicated that water quantity and quality in North America is particularly sensitive to climate change. Numerous studies have found that for a warmer and wetter climate, such as that expected for the Okanagan, there will be a decrease in the peak flow (Cohen et al., 2001; Loukas et al., 2002; Merritt et al., 2006; Morrison et al., 2002) and a shift in timing of streamflow (Leith and Whitfield, 1998; Merritt et al., 2006).

Cohen et al. (2004), predict that over the next 80 years temperature will increase in both summer and winter; and precipitation will increase in winter and decrease during periods of high demand in summer. Semi-arid regions, such as the Okanagan, are very sensitive to changes in precipitation. Winter precipitation will more likely fall as rain, decreasing the annual snowpack (Cohen et al., 2014). Taylor & Barton (2004) statistically downscaled six climate models and indicated an increase in temperature (1.5 to 4^{o}C), wetter winters and drier summers. Follow-up studies by Merritt & Alila (2004) and Merritt et al. (2006) suggest that there will be significant changes to the annual hydrograph (increasingly rainfall dominated); reduced snowpack, earlier onset of spring freshet, reduced spring freshet flow volumes, and decreased summer precipitation. Whitfield & Cannon (2000) showed that climate change has already caused a change in both the timing and magnitude of peak flows in the Okanagan and neighbouring Similkameen valley.

Surface water in perennial and intermittent streams is important contributors to the Okanagan River, either directly or through groundwater recharge. Smerdon et al. (2009) determined for BX creek that the majority (58%) of recharge to the valley aquifer occurs at the apex to the alluvial fan on the mountain side. Most recharge occurs post snow melt and often in March but is dependent on the snow pack depth and the timing of the melt (Smerdon et al., 2009). August is characterized by negative recharge, indicating evaporation from the groundwater. However, a recent isotope groundwater study suggested that there is no groundwater recharge from headwater streams (Wassenaar et al. 2011). Several other groundwater studies have been carried out in the Okanagan, including mapping fractures (McElhanney Consulting Services Ltd., 2006), mountain front recharge (Welch & Allen, 2012), modelling future recharge using climate change scenarios (Toews and Allen, 2009) and the importance of groundwater inputs and withdrawals on surface water during low flows (Obedkoff, 1990). The interactions between surface water and groundwater have been studied for a few reaches of the main channel of the Okanagan River and focused on salmon habitat (Neumann and Curtis, 2016). The Water Supply and Demand Project of the Okanagan Basin Water Board concluded that there is a lack of information about streamflow-groundwater interactions and understanding of flow regimes of small streams during low and no flow periods (Guy 2010).

Streams that have periods of no-flow (i.e. temporary streams) have been rarely studied in the Okanagan, even though research in other arid regions highlights the need to understand these systems, particularly as climate change will likely lead a larger number of streams that become intermittent for portions of the year. Huxter and van Meerveld (2012; Chapter 2) showed that there are differences in the stream flow regimes of intermittent and perennial streams but that more work is needed to understand intermittent stream flow regimes.

3.3.2. Study site: Long Joe Creek

Long Joe is a deeply incised, steep, first order stream and a typical temporary stream for the arid and semiarid Okanagan and southern Similkameen. Long Joe Creek is a typical southern Okanagan intermittent stream. It has periods of no flow every year and a comparable drainage area and slope to surrounding catchments. There are no artificial withdrawals from the catchment above Highway 3. Below Highway 3, a farm withdraws an unknown amount of water when there is streamflow. Large portions of the catchment were burned in the July 2003 Anarchist Mountain Fire. Long Joe Creek is used for range land for cattle in fall, which has caused numerous compacted areas along the hillslopes and minor alterations to the streambed. There are several deactivated forest service roads in the catchment and an incomplete housing development near the southern boundary of the catchment.

The catchment area of Long Joe Creek is 3.4 km^2. The main channel is 6.7 km long and has an average slope of 19% (Figure 3.1). The source of Long Joe creek is located at 1541 m above sea level (asl). From there it flows to Osoyoos Lake at 276 m asl. Annual average precipitation in Osoyoos is 323 mm/year, of which 44 mm (14%) falls as snow (Environment and Climate Change Canada, 2011). The presence or absence of ice on Osoyoos lake affects the cloud cover. During the 2010-2011 winter, the lake surface was ice covered from December to February. The winter lake effect regularly caused fog above 1000 m asl. Precipitation is fairly evenly distributed over the year, although it is slightly higher in May, June, and November. The annual average temperature is 10.4 ^{o}C, with summer (July-August) highs in the upper twenties degrees Celsius and winter lows (January-December) below zero degrees Celsius (Canada 2011). The 2011 study year was cooler (1.5 ^{o}C) and only slightly drier (0.2 mm) than the 1996-2011 average.

The geology (Figure 3.1) of the catchment consists of colluvium and glacial drift overlying fractured bedrock (Bowen et al. 2005; Roed and Fulton 2011). The bedrock at the headwaters of Long Joe consists of granodiorite, hornblende-biotite, and gneissic intrusive rocks known as Osoyoos Lake Gneiss (Late Triassic) and are overlain by surficial beach and dune deposits from the last glaciation (25,000 to 10,000 years ago). The lower slopes of the catchment consist of Osoyoos Lake Gneiss overlain by outwash terraces from the last glaciation. At approximately 800 m asl, the catchment narrows into a canyon comprised of greenschist metamorphic, rocks known as the Anarchist Schist (Permian and Triassic age). The upper catchment (above 1100 m asl) is dominated by

Anarchist Mountain Granite (early and middle Jurassic) bedrock overlain by colluvium of varying depths (Nasmith 1962; Okulitch 2013). Soil depths varies across the catchment; the major soil type is brown chernozem.

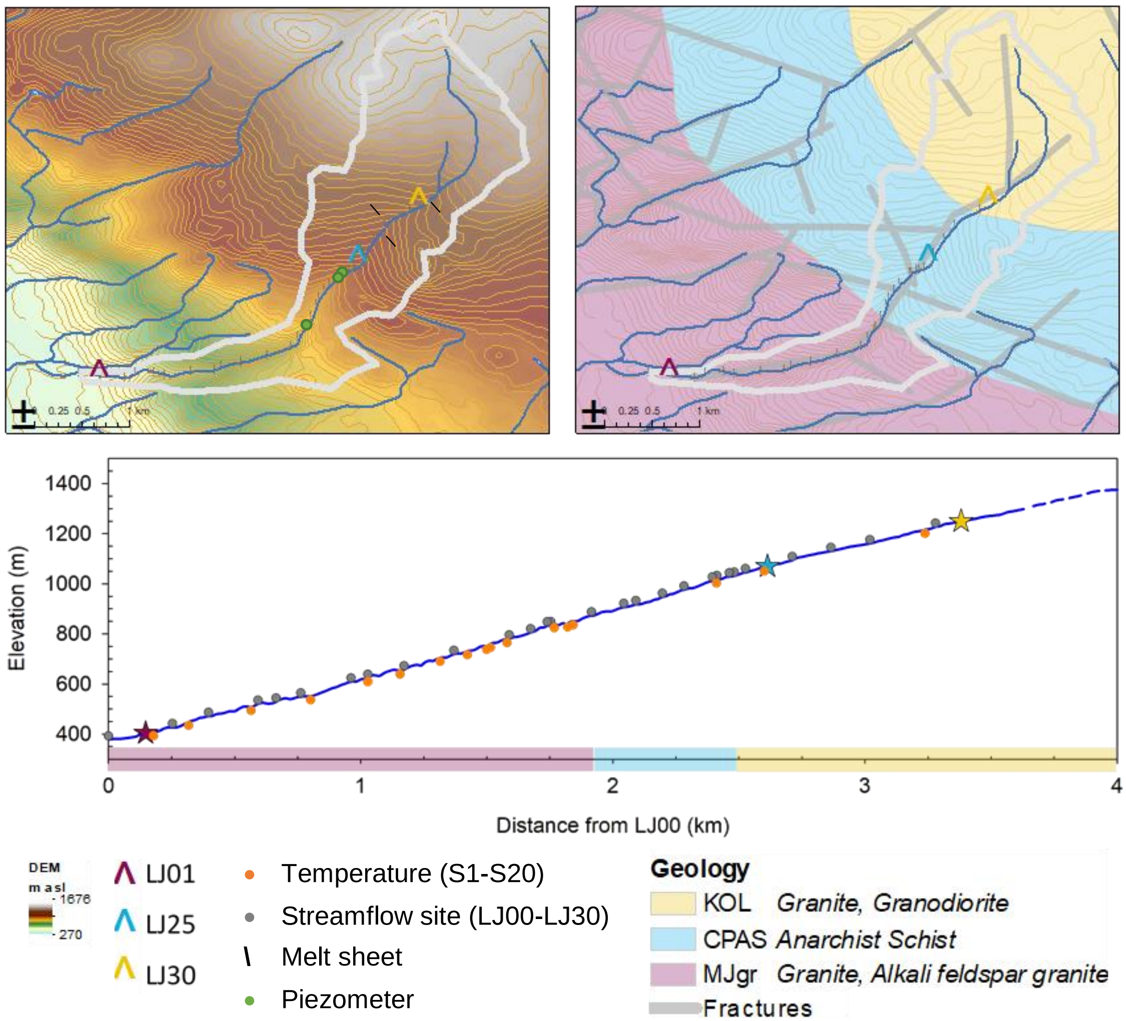

Figure 3.1 Long Joe Creek Catchment with the locations of gauging stations (LJ01, LJ25, LJ30), streamflow measurement sites (LJ00-LJ30), location of temperature probes (S1-S20), and geology.

Long Joe Creek is located in the southern Okanagan Bunchgrass Biogeoclimatic Zone with Antelope Brush Ponderosa Pine and Interior Douglas-fir zones. In the lower catchment (below ~850 m asl) the land cover is grasslands, composed of Bluebunch wheatgrass (Pseudoroegneria spicata), Idaho fescue (Festuca idahoensis Elmer) and rough fescue (Festuca scabrella), Sandberg's bluegrass (Poa secunda), junegrass (Koeleria macrantha K. cristata), yarrow (Achillea millefolium) and arrowleaf balsamroot (Balsamorhiza sagittata), Antelope brush (Purshia tridentate), snow buckwheat

(Eriogonum niveum), big sagebrush (Artemisia tridentate) and compact selaginella (Selaginella densa). Around 850 m asl the catchment vegetation transitions into open ponderosa pine (Pinus ponderosa) forest. Above 950 m asl the vegetation consists mainly of interior Douglas fir (Pseudotsuga menziesii), trembling aspen (Populus tremuloides), Lodgepole Pine (Pinus contorta), and Western Larch (Larix occidentalis). Vegetation in the riparian area consists predominantly saskatoon (Amelanchier alnifolia), common chokecherry (Prunus virginiana), Douglas maple (Acer glabrum) and mock-orange (Philadelphus lewisii). Horsetail is present in permanently wet locations.

3.4. Methods

3.4.1. Precipitation and snowmelt

Rainfall was measured at three locations in the catchment (at the lower and middle streamflow gauging stations (LJ01 and LJ25; Figure 3.2), and at a mid-slope location in the upper reaches of the catchment) using RG3 Onset HOBO tipping bucket rain gauges with a resolution of 0.2 mm/tip up to a maximum rainfall rate of 127 mm/hour. Each gauge was calibrated in the field prior to installation and during the data collection period. An average calibration value was used for the precipitation calculations. Additional rainfall data was obtained from the Environment Canada Osoyoos weather station (ID: 1125852), located 1.6 km from the bottom gauging station (Environment Canada Daily Precipitation, 2013) and from the B.C. Ministry of Highways and Infrastructure Anarchist Summit weather station (ID: 33093) located on the north side of Highway 3 in a gravel pit 7.5 km south of Long Joe (Province of British Columbia, 2013). The Anarchist Summit station is not maintained in the summer months.

Snow water equivalent (SWE) measurements were taken on eight 50 m long transects (Figure 3.2) on March 11-13, April 1-2, April 30th and May 5th 2011, using a Federal snow sampling tube. Snow depth data were also obtained from the B.C. Ministry of Highways and Infrastructure Anarchist Summit weather station (33093). Melt water was collected from three melt sheets connected to RG3 Onset HOBO tipping bucket rain gauges to collect data on the timing and relative rate of snowmelt.

3.4.2. Gauging stations

Water level and Electrical Conductivity (EC) were measured at five minute intervals at three locations in Long Joe (Figure 3.8) using OTT PLS pressure transducers and Campbell Scientific CS547A-L conductivity probes connected to CR1000 loggers. The lower gauging station (LJ01) is located 1.8 km from the lake at 405 m asl (100 m from Highway 3 and an orchard), the middle gauging station (LJ25) is located just above the fog line and the canyon, and the upper gauging station (LJ30) is located 1.1 km below the source of Long Joe Creek in the flatter upper area north of Anarchist Mountain (Figure 3.1). The accumulated area of each station is approximately twice the area of the upstream station (LJ30 1.0 km^2, LJ25 2.0 km^2, LJ01 3.6 km^2). Water levels were measured relative to an arbitrary datum at each site. Streamflow was measured at all three locations using dilution gauging with fluorescent dye (Rhodamine WT (Water Tracing)). Rating curves were developed based on 20 discharge measurements at each site. The measurements plot within 5% of the rating or 3 mm. On May 26th 2011, a large rain event occurred, which caused numerous landslides in the catchment and destroyed the mid-station (LJ25). The station was re-established on June 8th 2011.

A comparison of the hydrographs for the upper (LJ30), middle (LJ25) and lower (LJ01) gauging stations allows for the determination of periods of gaining and losing stream conditions and the time of drying. If streamflow (l/s) at the mid station was more than 10% higher than the flow at the upper station or if flow at the lower station was more than 10% higher than flow at the mid-station, the stream reach between the stations was considered to be a gaining reach. Similarly, if the decreases in flow were larger than 10%, it was considered to be a losing stream. Differences within 10% were considered to be within the error margin and indicate neither gaining nor losing conditions.

3.4.3. Streamflow measurements along the stream channel

The entire stream length was surveyed and the occurrence of flow/no flow was mapped on February 13th to 17th (but streamflow was not measured at this time), August 17th and 18th, and September 25th and 26th, 2011. The February mapping was indicative of mid-winter flow patterns, while the August and September flow mapping provided data during the period of streamflow cessation.

Streamflow was measured 21 times between May 3rd and October 21st, 2011 at an additional 28 locations (in total 31 stations LJ00-LJ30) along the stream (Figure 3.1;

Table 3.1) using dilution gauging with Rhodamine WT (RWT) and a Turner Design Cyclops-7 submersible probe. Two days were required to take streamflow measurements at all 31 locations. In order to be able to compare the measurements on the two days, the last measurement on the first day was repeated at the start of the second day. The volume of RWT depended on the water level and predicted streamflow for the day. The Cyclop-7 probe was calibrated every day for the RWT solution used.

To account for the diurnal variation in streamflow, streamflow measurements during the longitudinal surveys were converted to discharge at a fixed time (18:00) when there was no rainfall. A sine wave was fitted to the data from the three gauging stations (LJ01, LJ25 and LJ30) and the lag time and amplitude were correlated to the elevation. A sine wave with an amplitude and lag time for the station (based on this correlation for the 3 gauging stations) was used to transform the measurement data to an estimated discharge at 18:00. This allowed for comparison of discharge at the different stations and the determination of changes in total discharge over the entire length of the stream, even though the measurements were not taken at the same time. Notes on the occurrence of flow were taken each time as well. The contributing area to each station was determined to determine how discharge (streamflow per unit area) varied throughout the catchment. At all 31 streamflow locations sediment size distribution was measured using Wolman counts, channel slope and channel cross sectional area were measured. Bankfull flow was calculated for each site using Manning's equation (Table 3.1).

$$Q = vA = \left(\frac{1}{n}\right) AR^{\frac{2}{3}}\sqrt{S}$$

Q = discharge (m^3/s)

v = velocity (m/s)

A = cross sectional area (m^2)

n = Manning's roughness coefficient

R = hydraulic radius (m)

S = channel slope (m/m)

Table 3.1 **Streamflow measurement sites and stream reach characteristics. n = Mannings roughness coefficient. Station LJ01 is the lower station, LJ25 is the mid station and LJ30 is the upper station.**

Station	Distance from bottom (m)	Accumulated Area (km^2)	Occurrence of No Flow	Pebble count data		Slope		Bankfull flow (l/s)	
				D50 (m)	D84 (m)	(deg)	(m/m)	*n*=0.007	*n* based on D84*
LJ00	0	3.62	YES	0.155	0.550	7.4	0.13	781	198
LJ01	175	3.60	YES	0.054	0.180	7.4	0.13	673	626
LJ02	331	3.58	Pooling			14.5	0.26	792	NA
LJ03	423	3.57		0.026	0.099	6.1	0.11	882	1299
LJ04	557	3.53		0.073	0.310	8.5	0.15	829	548
LJ05	635	3.50	YES	0.023	0.117	7.8	0.14	288	288
LJ06	823	3.43		0.110	0.330	11.9	0.21	761	457
LJ07	978	3.35		0.122	0.530	18.9	0.34	805	144
LJ08	1038	3.33	YES	0.122	0.900	18.5	0.33	802	2
LJ09	1183	3.26	YES	0.240	0.471	18.4	0.33	720	164
LJ10	1378	3.02	YES	0.083	0.380	15.8	0.28	693	321
LJ11	1562	2.85	YES	0.300	1.548	19.1	0.35	762	0
LJ12	1663	2.82		0.047	0.360	11.5	0.20	603	227
LJ13	1761	2.79	YES	0.180	1.400	17.2	0.31	1613	0
LJ14	1780	2.78	YES	0.120	0.710	17.7	0.32	459	0
LJ15	1899	2.73	YES	0.114	0.410	12.4	0.22	856	347
LJ16	2022	2.70	YES	0.060	0.140	10.7	0.19	711	804
LJ17	2126	2.66	YES	0.053	0.154	5.6	0.10	638	513
LJ18	2196	2.58				19	0.34	836	NA
LJ19	2270	2.57	YES	0.058	0.290	19.3	0.35	874	500
LJ20	2409	2.16	YES	0.168	0.530	24.6	0.46	2688	1064
LJ21	2410	2.16	YES	0.240	0.620	27.8	0.53	210	0
LJ22	2461	2.12	YES	0.048	0.161	9.6	0.17	83	36
LJ23	2462	2.12	YES	0.044	0.186	4.2	0.07	1068	1070
LJ25	2592	2.02		0.102	0.305	10.9	0.19	794	245
LJ26	2679	1.94		0.092	0.324	9.6	0.17	932	602
LJ27	2848	1.66		0.032	0.180	9.4	0.17	755	636
LJ28	3006	1.56		0.080	0.478	12	0.21	679	146
LJ29	3243	1.10	YES	0.146	0.333	14.3	0.26	575	189
LJ30	3349	1.05	YES	0.123	0.291	7.7	0.13	311	91

$$^{*}n = \frac{0.1129 \cdot R^{\frac{1}{6}}}{1.16 + 2 \cdot \log\left(\frac{R}{D84}\right)}$$

3.4.4. Groundwater level and EC measurements

At the upper (LJ30) and lower gauging station (LJ01) groundwater piezometric head was measured in 5.8 cm diameter piezometers with the screen (10 cm screened) located 50 cm below the ground surface using an OTT PLS pressure transducers. These readings were compared to the stream level to determine whether the stream was gaining of losing. Shallow soils at the mid-station prevented the installation of a piezometer there. Additional groundwater water levels were collected manually at three other piezometers located in the watershed (Figure 3.1): one piezometer was located 24 m upstream of LJ22 next to the main channel in an area that remained snow free in the winter, another piezometer was located 2 m downstream of LJ22 at the top of a large drop in streambed elevation, and the third piezometer was located in a horsetail patch between LJ13 and LJ12.

The Electrical Conductivity (EC) of the water was measured to determine the source of the stream water at different locations. The stream EC was measured continuously at LJ01, LJ25 and LJ30, and groundwater EC at LJ01 and LJ30 using a Campbell Scientific CS547A-L conductivity probe. Additionally, stream EC measurements were taken at all 31 streamflow measurement locations, and at a few seeps and areas of surface flow throughout the catchment using a portable Hanna Instruments 991300 probe. There are no EC data for LJ25 from July 3rd to 17th and August 25th to September 11th because the EC probe was out of the water. At LJ30, data are missing for the period September 6th to 26th due to power failure.

3.4.5. Temperature measurements

Streambed temperature data can be used to determine the timing and spatial patterns in streamflow and the general water source. Onset TidbiT v2 temperature probes were installed throughout the catchment in February 2011. The probes were shaded and attached to the streambed in locations of flow or no flow, roughly 80 m apart or where there were transitions from flow to no flow in February (Figure 3.1). The probes were installed in late February 2011 and started recording data on March 1st until late fall. In total 42 probes were installed but only 20 had data at the end of the field season (S1 to S20; S1 is the lowest elevation probe, S20 is the highest elevation probe). S17 was lost on May 26 and replaced by S19 on June 8th; S18 was installed June 8th. Additionally, temperature was measured at three profiles along the stream (Figure 3.1) at LJ01 (probe depths: surface, 0.15 m, 0.30 m, and 0.45 m), LJ22 upper located 24m upstream of LJ22 (probe depths: 0.01 m and 0.38 m) and LJ22 lower, 2 meters downstream of LJ22 (probe depths: surface, 0.03 m, 0.20 m, 0.35 m and 0.60 m). Air temperature was measured at 9 locations with 6 having data at the end of the field season (A1 to A6). Air temperature was also measured at each of the three gauging stations using a shielded Campbell Scientific 107-L temperature sensor.

Data from these probes was plotted to determine the changes in diurnal variations and when and where in catchment streamflow was present. The ratio of the standard deviation of stream temperature and air temperature was also calculated for each probe daily using the air temperature data from the probe located nearest to the stream probe.

3.5. Results

3.5.1. Rainfall during the study period

Precipitation data from the Environment and Climate Change Canada Osoyoos weather station (Osoyoos CS 1125852), LJ01, LJ25 and the Ministry of Transportation and Infrastructure weather station (33093 ANARCHIST SUMMIT) were used to determine the variation in precipitation over the catchment. Precipitation increased with elevation. The majority of rainfall events were recorded at all weather stations and likely fell over the entire watershed (Figure 3.2B).

Spring 2011 was cold and wet in the Okanagan. The average maximum temperature at the Osoyoos station between March 1 and May 31 was 2.5^{o}C lower than normal (2011: 14.9 ^{o}C; 1996-2011: 17.4 ^{o}C), while the mean temperature was 1.3 ^{o}C lower than normal (2011: 9.3 ^{o}C; 1996-2011: 10.6 ^{o}C). Precipitation in spring (March 1 to May 31) was 40.2 mm higher than normal (2011: 130 mm; 1996-2011: 90 mm). Reports by farmers indicate that fruit trees in the valley bottom bloomed approximately 3.5 weeks later than normal. Late summer and fall (Aug-Oct) 2011 were drier (2011: 26.6 mm; 1996-2011: 62.8 mm) and warmer than average (2011: 17.1 ^{o}C, 1996-2011: 16.5 ^{o}C).

Snowmelt started to increase on March 18th; with the snow pack melting rapidly so that the mid elevation gauging station was snow free by the end of March (Figure 3.2). Snowfall at the mid-station elevation after this day melted within 24 hours. Snow was present on the ground at the upper gauging station (LJ30) until May 6th. The last snowfall in the watershed occurred on April 28th, when the snowline was 200 meters below LJ25, approximately at the location of the valley fog line in the winter.

During the study period, there were five periods of higher rainfall: mid-March, May, early June, late July, and early October (Figure 3.2B). The March events coincided with the initiation of the spring melt, the May precipitation events led to the peak discharge for the year and the early October events ended the period of no flow in the stream (Figure 3.2 B and C). Post freshet, convective storms were the main source of precipitation.

A fog layer is present throughout the south Okanagan during the winter months, especially if Osoyoos Lake does not freeze over. The fog line is usually located at about 1000 m asl and varies in thickness, often the upper reaches of the mountains are above the fog line. Stations upstream of LJ19 were often in the fog during the winter. This fog layer reduced solar radiation in the mid-and upper reaches of the catchment, therefore helping to maintain the snowpack and created high levels of moisture in the air and in the ground.

3.5.2. Streamflow at the upper (LJ30), middle (LJ25) and lower (LJ01) gauging stations

Frozen stream conditions existed until April 23rd at LJ30, and until April 2nd at LJ25. LJ01 was not affected by ice during the period of study. The increased melt from the lower to mid-elevations of the catchment was detectable in the hydrograph of LJ01, i.e. the sharp rise in mid-March (Figure 3.2C). The discharge at LJ25 and LJ30 remained relatively constant until early May (Figure 3.2D), suggesting that melt water contributions to the upper parts of the stream were limited during this time.

During the pre-freshet period (i.e. prior to the upper part of the catchment being snow free), the stream in the upper portion of the watershed (LJ30-LJ25) was a gaining stream, while the lower portion of the stream (LJ25-LJ01) was a losing stream. The hydrographs from April 27th to May 1st demonstrate different processes within the watershed. Discharge at LJ25 and LJ30 changed relatively little and remained stable, while discharge at LJ01 increased over the entire period, but losing conditions remained at the lower stream section. This indicates that water came from upper areas and there were stream losses. On May 3rd there was a transition from losing to gaining stream conditions in the lower portion of the watershed. This suggests that streambed storage was filled, and/or stream flow was significantly higher than the streambed infiltration capacity and water was now carried down the channel. During this transition period (May 2-3) rain (3.4 mm in two days) and high night time temperatures may have caused significantly increased inputs (8 l/s to 19 l/s) of melt water to the channel.

From May 4th to 6th the lower stream reaches continued to be a gaining stream, although not as much as would be expected based on the contributing area. On May 5th at 10:00, flow at LJ25 was 81% more than flow at LJ30 (the drainage area doubles between LJ30 to LJ25), while flow at LJ01 was only 10% more than flow at LJ25 (drainage area increases 180%). This indicates that streamflow contributions from lower parts of the catchment were small or parts of the streamflow gains were offset by losses to storage/subsurface flow in the lower portions of the watershed.

Freshet streamflow was primarily a result of several rain events, melt due to higher temperatures and saturated soil conditions. In early May, the stream in the upper catchment tended to be a gaining stream, while the stream in the lower catchment was either losing or had very small gains. This changed on May 11th. The peak flows at LJ30 occurred on May 11th at 23:15 (303 l/s), at LJ25 on May 12th at 00:25 at (444 l/s) and at LJ01 on May 16th at 6:50 (864 l/s). LJ01 had a maximum flow of 428 l/s at 2:10 on May 12th, the day of the peak at the mid station. This suggests there were small gains or large gains that were offset by significant losses during transmission in mid May. However, the hydrographs during the freshet have a large uncertainty due to extrapolation beyond the measurements to establish the rating curve (extrapolation beyond 200%). For example, for LJ01 the highest measured discharge was 232 l/s, but the estimated peak discharge was 864 l/s.

Between May 12th and May 17th, daily minimum and maximum flows had different trends. Daily minimum streamflow increased between LJ30, LJ25 and LJ01. However, the daily maximum flows increased between LJ30 to LJ01 but there were losses at LJ25. Rain events on May 15th (Osoyoos 1.6 mm; Anarchist 2.5 mm) and 16th (Osoyoos 11.2 mm; Anarchist 12.0 mm) highlight the losses at LJ25. During the peak periods of these rain events the upper catchment was losing and the lower catchment was gaining streamflow. For example, peak flows on May 16th did not increase downstream:

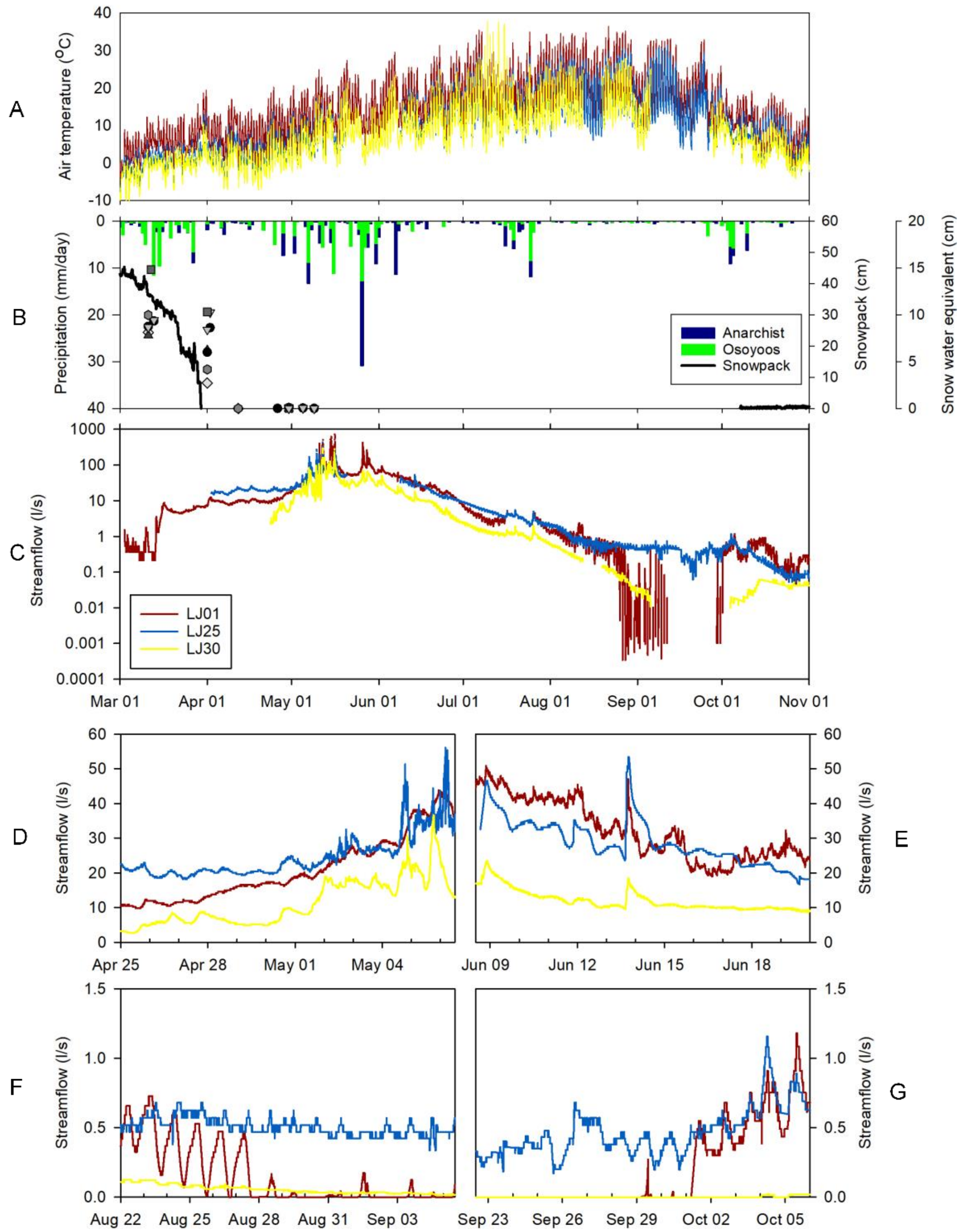

Figure 3.2: **Weather, snow and streamflow. A – Air temperature Osoyoos (red), LJ25 (blue), and Anarchist (yellow); B – Snowpack depths, SWE, and precipitation; C – Streamflow LJ01 (red), LJ25 (blue) and LJ30 (yellow); D April Streamflow at LJ01, LJ25 and LJ30; E – June Streamflow at LJ01, LJ25 and LJ30; F – August Streamflow at LJ01, LJ25 and LJ30; G – September Streamflow at LJ01, LJ25 and LJ30.**

LJ30 peaked at 302 l/s at 5:50, LJ25 peaked at 255 l/s at 8:15 and LJ01 peaked at 864 l/s at 6:50. At 6:20, when the peak flow should have arrived at LJ25, the discharge was only 61 l/s. A similar response was seen on May 14th and 15th. Between May 17th and 26th there was no rain and the stream was gaining from LJ30 to LJ25 to LJ01. On May 26th, a large rain event occurred (Osoyoos 12.8 mm; Anarchist 30.8 mm) causing numerous landslides and debris flows in the catchment, destroying the monitoring station at LJ25. During this event, flow at LJ30 increased by 32 l/s, while flow at LJ01 increased by 343 l/s. This large increase at LJ01 and lower rainfall in the valley indicates that there were fewer losses during transmission during this event compared to the peak flows in early May.

In early June, the peak of the freshet had passed and the watershed was drying. A small channel to the north of LJ01 dried out on June 6^{th} and a 40 cm deep hole near LJ04 dried on June 9^{th}. The valley bottom (Osoyoos) received 6 mm of rain in early June, while the upper mountains (Anarchist) received 24 mm of rain. Streamflow increased over the length of the stream during this period. On June 10^{th} at 10:00 am LJ25 had the same discharge as it had at 10:00 am on May 5^{th} (May 5^{th} and June 10^{th} were chosen for comparison as the mid-station has a similar discharge and these days represent patterns that are similar for the other days), but the percent increase from LJ30 to LJ25 was 152%, and the percent increase from LJ25 to LJ01 was 19%, i.e., double from the May 5^{th} situation. This indicates that there were larger contributions to streamflow from the lower portions of the watershed (and smaller losses to the streambed) and that groundwater was a larger contributor to streamflow in the lower part of the catchment in early June than May. Most of the soils in the upper part of the watershed (above LJ25) were still saturated and overland flow was seen on many of the banks near the stream, which could explain the larger increases in flow between LJ30 and LJ25.

Three significant rainfall events occurred between June 7^{th} and June 20^{th}, with very different effects at each of the three gauging stations. Between June 7^{th} and 20^{th}, the hydrographs at the three stations were largely on the recession. On June 7, 11.4 mm of rain fell between 17:00 and 20:00 at the Anarchist summit (no rainfall was recorded at Osoyoos; at Oliver there was 8.2 mm of rain), causing the peak in the hydrograph on June 8^{th}. Throughout this rain event, the stream remained a gaining stream. LJ30 and LJ25 had a very fast rising limb and falling limb, but LJ01 had a small increase at the peak of the event and then a very long recession. This muting of the hydrograph downstream is expected as a result of streamflow routing and a lack of rain in the lower part of the watershed but is not consistent with other events and may indicate that groundwater/interflow from upper portions of the watershed increased flow. The small rain event on June 13th (15:30-18:00, 1.8 mm at Anarchist, 2.2 mm in Osoyoos) had a very different effect on streamflow. The hydrograph of all three stations had an abrupt rising and falling limb and there was no muting of the hydrograph at LJ01. The stream became a losing stream in the lower portions of the watershed during the peak of the event. The rain started at 15:30 (from notes), the rising limbs of LJ25 and LJ30 started at 15:40, while flow at LJ01 didn't start to rise until 16:25. This short response time could suggest that the rise at the upper station was largely due to rain falling directly on the channel and overland flow in the saturated upper areas of the watershed. While the whole watershed received rain at similar times, in lower drier areas more of the rain was lost to evaporation or stored in the ground, not making it to the stream, and the delayed response of the hydrograph at LJ01 was likely water from the upper watershed making it down the channel to LJ01. Due to the dryness of the lower watershed, there were losses to the streambed and streamflow at LJ01 was less than at LJ25. Additionally, the rainfall amount was not enough to cause a larger increase in interflow in the watershed, which likely occurred during the June 7th storm. However, the response during the very small rain event on June 18th (1.7 mm at Anarchist; 0 mm in Osoyoos, but field notes say light rain even in the valley bottom) does not fit this theory. There was almost no change in streamflow at LJ25 or LJ30, while at LJ01 there was a marked increase in streamflow over a long period (from mid-day on the 17th to the 20th the lower portion of the catchment is gaining).

The diurnal variation in streamflow increased as temperatures increased and streamflow became less. By late June, diurnal variations in streamflow were prominent at all three stations, although the amplitude decreased with elevation, which is consistent with smaller temperature changes at higher elevation, and/or less evapotranspiration.

In early July, the lower catchment was a losing stream (LJ01 had less flow than LJ25); this is likely explained by losses to evapotranspiration being greater in the lower portions of the catchment. LJ30 consistently had the lowest discharge, as would be expected based on its small contributing area. Between July 16-19, there were numerous rain events. The reduced evapotranspiration and precipitation cause LJ01 to have the highest flow again but three days after the July 25th rain event, the lower catchment (between LJ25 and LJ01) was again a losing stream section.

Sustained high temperatures in late July and early August caused continued drying of the catchment. The mean streamflow at LJ01 and LJ25 differed by only 0.15 l/s , with particularly large diurnal variations at LJ01. Towards the end of August, diurnal variations in streamflow disappeared at LJ25, even though the temperature was high, while at LJ01 the diurnal signal became large and the streamflow became zero every day. This could indicate the water at LJ25 was primarily groundwater and that channel evaporation rates and channel infiltration were greater than streamflow, causing the channel to dry between LJ25 and LJ01.

On September 11th, LJ01 was dry and LJ30 had no flow but a damp bed, while at LJ25 there was continuous flow that was relatively stable at 0.5 l/s. This supports the earlier inference that groundwater drives summer discharge at LJ25. The streambed at LJ30 remained damp but there was no flow, indicating that the groundwater table was near the surface or subsurface flows contributions existed but were not a large contribution to summer flows. On September 16th, streamflow at LJ25 decreased from 0.4 l/s to 0.2 l/s during the day, streamflow continued to decline afterwards, with a larger diurnal signal. This could suggest that the majority of the groundwater input was from storage from freshet and no longer available.

The September 26th rain event resulted in an increase in streamflow at LJ25; a day later flow returned to LJ01. The October 1st rain event resulted in LJ01 having flow of similar magnitude as the flow at LJ25; the lower stream reach was neither gaining nor losing. Rainfall continued until October 5th. At LJ30, flows were intermittent on the 4th and permanent by the 5th of October. All three stations had recessions until the next rain event (Oct 10th). After the peak of this event, streamflow increased from LJ30 to LJ01; the stream remained a gaining stream for the duration of open water (i.e. until some parts froze later in winter).

3.5.3. Spatial variation in streamflow along the stream channel

Streamflow measurements at 31 stations along the stream channel (LJ00 to LJ30) during the spring-fall season complement the inferences on the gaining and losing character of the stream reaches from the gauging station data. Figure 3.3 shows the change in streamflow along the stream for six of the 21 measurement days. These six days were chosen because they represent the seasonal variation in streamflow.

Long Joe had gaining and losing sections along the stream length throughout the year (Figure 3.3 and Figure 3.4). Overall the upper catchment was predominately a gaining section, with the majority of the gains occurring between LJ28 and LJ25, until streamflow was almost non-existent throughout the catchment. Streamflow losses in the lower catchment occurred most consistently at LJ01, LJ05, LJ08, LJ11, LJ14 and LJ18 (Figure 3.3 and Figure 3.4). Streamflow most consistently increased at LJ03, LJ04, LJ07, LJ10, LJ12, LJ13 and LJ19 (Figure 3.3 and Figure 3.4). As would be expected the gains decreased over the season and the losses increased. Also, the total number of gaining reaches decreased and the number of losing reaches increased.

In early May. Long Joe was generally gaining streamflow in the upper portion of the watershed above LJ25, while there were only small increases in flow from LJ19 to LJ00. The losses were greater than 10% at LJ11 and LJ14 (LJ08 -7.5%) (Figure 3.4), while the gains were greater than 10% at LJ07, LJ13, LJ19. Otherwise below LJ19 the gains and losses were less than 5% and thus less than the assumed measurement accuracy.

The pattern in late May was similar, with most of the gains in the upper catchment, while the lower stations had small gains or losses. The pattern of loss (~5%) at LJ08, LJ11 and LJ14 were still present but were not as significant as in early May. Streamflow decreases between LJ00 and LJ01 were ~5% in early May but during peak freshet these were gaining stream segments. Similarly, the stream right above LJ25 was a losing stream segment pre-freshet peak, but it became a gaining section during freshet (Figure 3.4).

In June, the majority of the gains also occurred in the upper portion of the catchment; below LJ19 the gains only just barely outweighed the losses. Post peak freshet, LJ02 became a major gaining stream segment but this changed by the end of June as the catchment started to dry out. The largest gains in the upper catchment occurred in the sections between LJ28 to LJ26 and at LJ19. The relative gains at the other stations tended to be less than 5%. For LJ07, LJ13 and LJ19 the relative streamflow gains increased as the catchment dried out. During freshet the relative gains were insignificant, by mid-June they were larger than 10% and towards the end of June, the relative gains were close to 20%. On June 15th, the first loss greater than 10% occurred at LJ11 (although for LJ08 and LJ14 the losses were just under 10%). As June progressed, there was a noticeable increase in losses at LJ18 and LJ22/23. LJ18 is located on bedrock.

July 15th was the first time that the minimum discharge in the catchment was not at LJ30, but instead was at LJ08. In July, there were fewer gaining streamflow sections at the very top of the catchment and by the end of July there were losses at LJ29. LJ28 to LJ26 continued to be an area of relatively large gains in streamflow. Downstream of LJ25 the gains were becoming smaller and smaller. Although LJ07, LJ13 still had large relative gains, LJ19 was no longer a gaining station. LJ17 started to have significant losses in streamflow. LJ02 had very small relative gains in June, yet in July there were significant gains. LJ01 and LJ00 continued to be losing sections, as they did in June. LJ08, LJ11 and LJ14 continued to be losing sections as well, as they had for the past two months. LJ05 and LJ09 started to have more significant losses. By the end of July, LJ20 to LJ23 also had large losses.

August 17th was the first survey during which zero flow was observed at several location (Figure 3.5), although it likely happened before this date as well. Zero flow was recorded at LJ00, LJ8 to LJ11, LJ14, LJ16, LJ17 and LJ20-LJ23. LJ02 was a losing section again, while LJ03 was now a gaining section. The section between LJ28 and LJ25 was still a gaining section. Downstream of LJ25 (until LJ05), anywhere there was flow it was similar (~0.55 l/s, except for LJ19 where it was lower).

Throughout July, August and September the station with the maximum flow moved upstream (from LJ02, LJ05, LJ06 to LJ26). By the end of September, there was very little flow in the entire catchment and the only stations with flow were LJ03, LJ04, LJ06, LJ07, LJ12, LJ26 to LJ28 (Figure 3.3 and Figure 3.5). Interestingly LJ13 went dry even though it was a consistent gaining section up to this point, suggesting there were no longer groundwater inputs to this section. Of the lower stations, LJ06 had the highest flows, which was about half of the flow at LJ26.

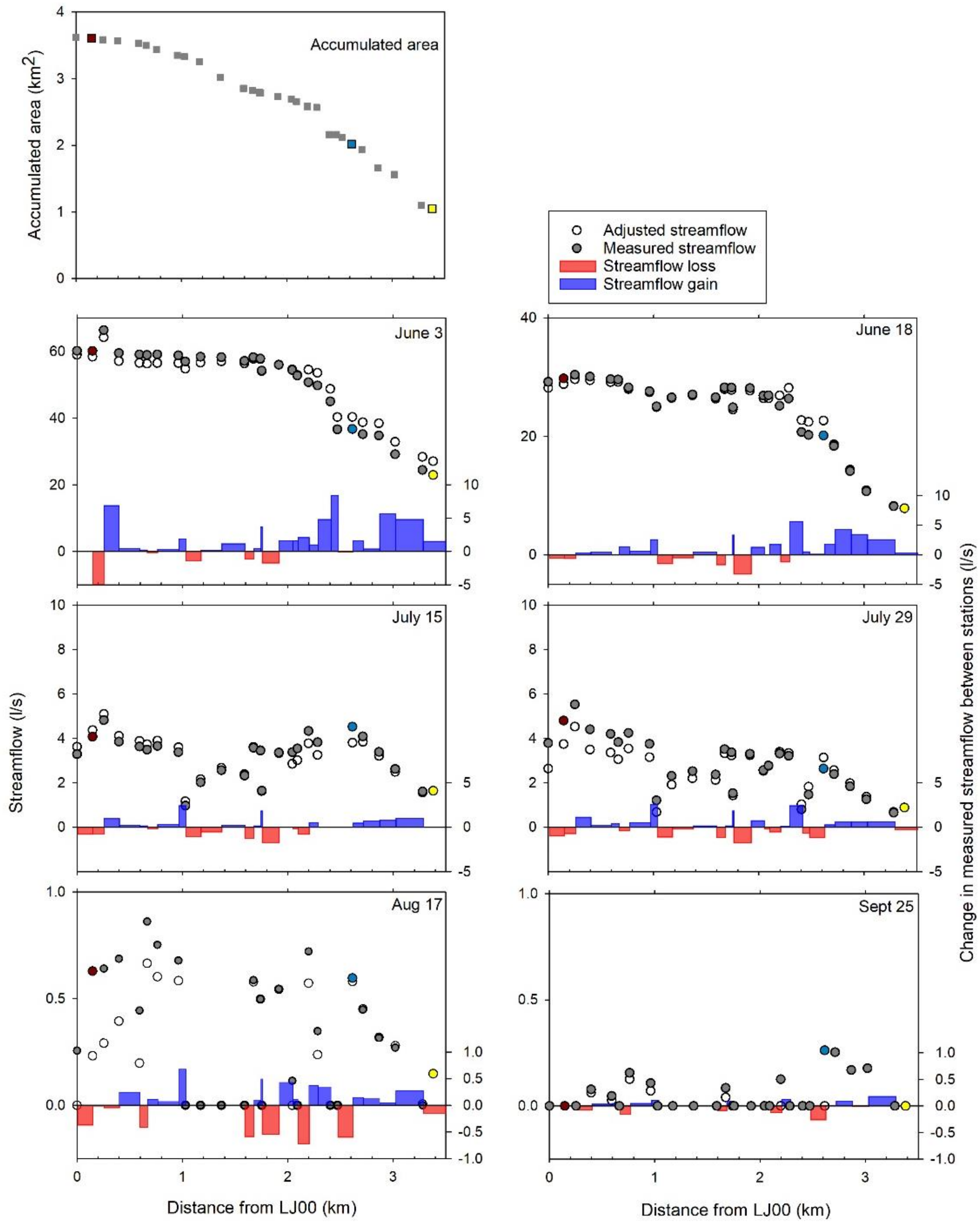

Figure 3.3: Spatial variation in streamflow along the stream channel (LJ00 to LJ30). The solid back points represent the actual streamflow measurement, the open circles represent the estimated streamflow at 18:00 (based on the fitted sine-wave), red bars indicate that there is a loss in streamflow from the preceding station and a blue bar indicates there is a gain (based on the actual discharge measurements). Coloured dots represent the continuous water level stations (maroon LJ01, blue LJ25 and yellow LJ30).

The occurrence of flow along the channel in late summer was highly variable (Figure 3.5). For example on September 25, there were 14 transitions from flowing to non-flowing conditions along the channel (Figure 3.6). There were longer sections with and without flow but there was no clear relation between the occurrence of flow and topography. For example, the reach slope of each station could not predict whether the reach went dry or not; some very steep reaches always had flow but other low angle reaches also always had flow. All streamflow measurement stations with coarse bed

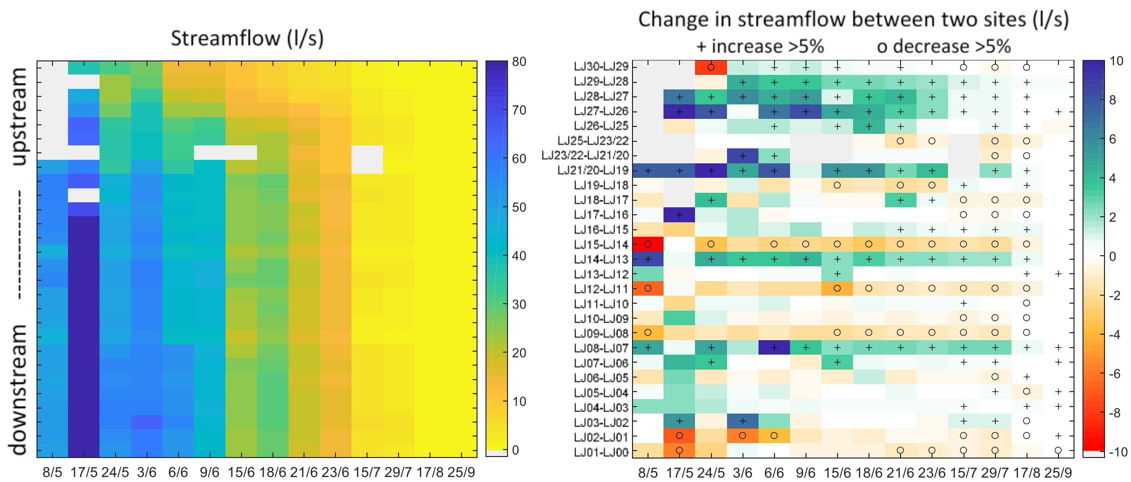

Figure 3.4 **Streamflow at each station and change in streamflow from upstream to downstream station for different measurement dates**

material went dry during the summer but fine bed material did not guarantee permanent flow (Figure 3.7). LJ18 and LJ02 were excluded from these analyses because both were bedrock. Other measured stream characteristics did not appear to provide any insight as to whether a section would go dry or maintain streamflow. High or low bankfull flows (calculated using an estimated roughness (*n*) value of 0.007 and using Limerinos (1970) calculated *n* based on the D84 value) were not indicative of dry or flowing reaches either.

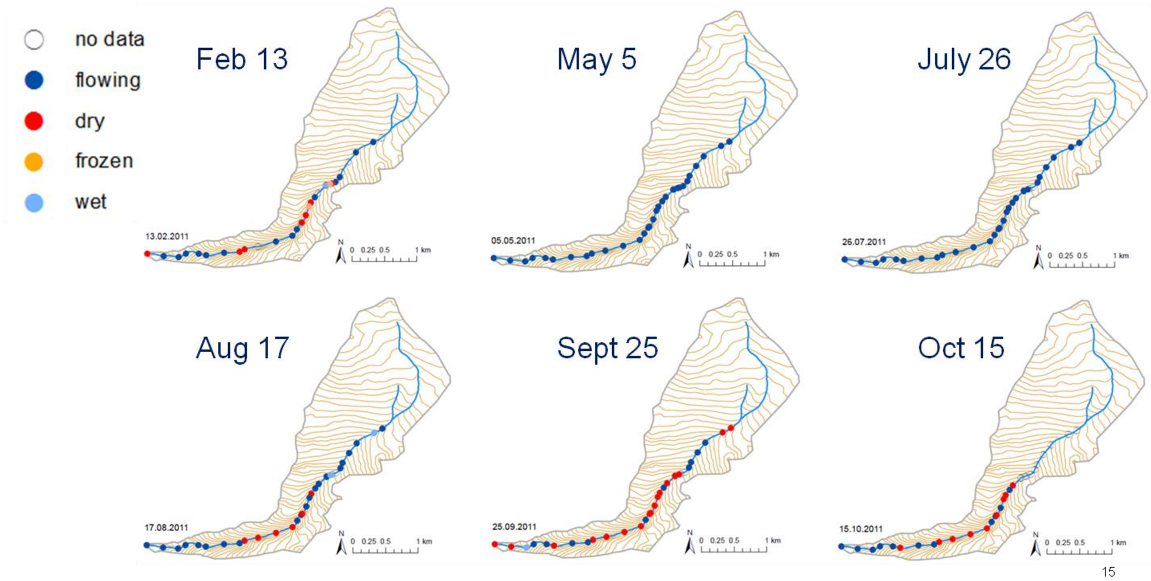

Figure 3.5 **Occurrence of flow LJ00 to LJ30 for 6 time periods. Dots are locations of each station LJ00 to LJ30. Dot colour indicates flow (dark blue), no flow (dry) or pooling water (light blue).**

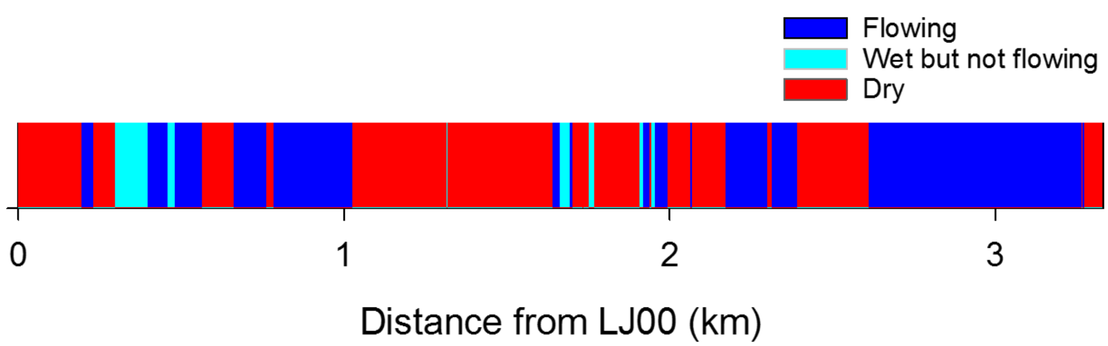

Figure 3.6 **Occurrence of flow, wet channel beds and dry channel beds along long Joe on September 25, 2011 based on detailed mapping of the entire stream.**

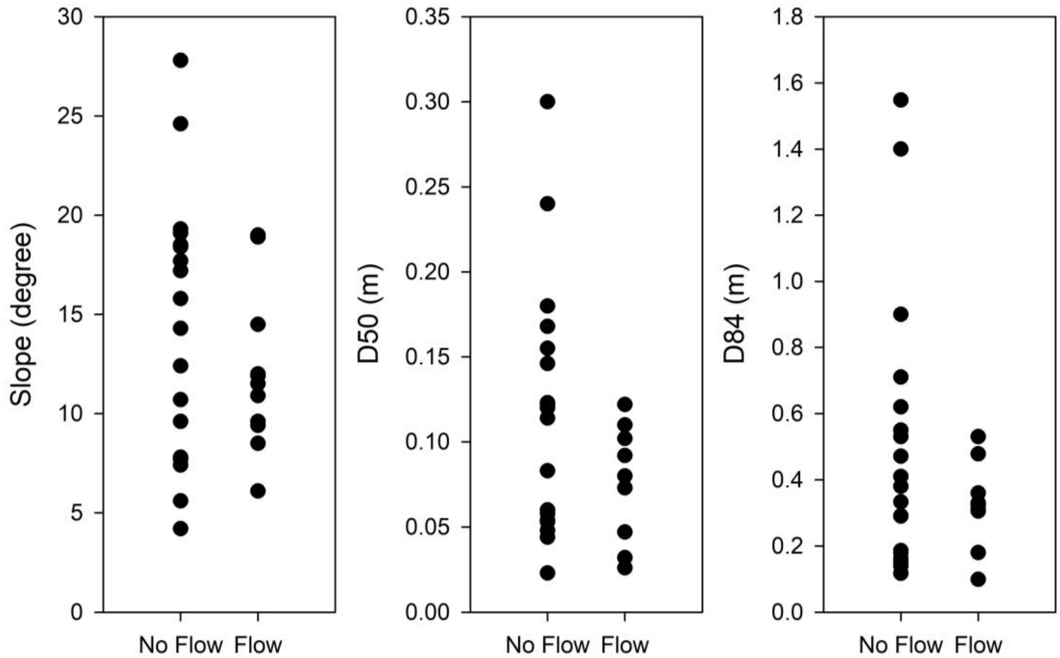

Figure 3.7 Slope, D50 and D84 at stations with flow or no flow.

3.5.4. Electrical conductivity

Three gauging stations

The electrical conductivity in streamwater increased from the headwaters to the outlet (Figure 3.8). Over the entire season there was an inverse relationship between conductivity and streamflow. Pre-freshet, LJ25 and LJ30 had a stable EC at 80 μs/cm and 45 μs/cm respectively. EC at LJ01 increased from 100 μs/cm in mid-February to 158 μs/cm on March 28, and then began a decline until the peak of the freshet. From mid-March onward, there was a clear diurnal pattern in the EC at LJ01. The decline in EC at LJ01 was not smooth, and the steps did not coincide with rain events. EC at LJ25 started to decline in early April with a diurnal pattern starting on April 8th . By early May it decreased rapidly indicating that the melt water fraction in the stream increased. At LJ30, EC remained stable until early May, with little to no diurnal pattern and then began to decline. The timing of the decline in EC and the occurrence of diurnal patterns coincided with significant snowmelt at each station.

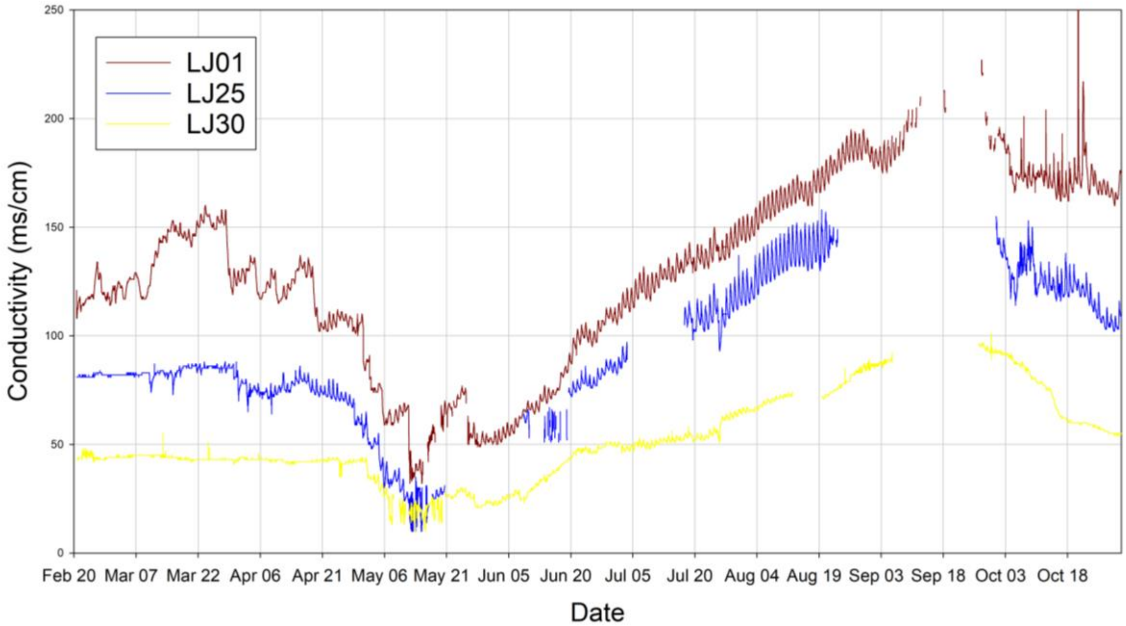

Figure 3.8 Stream electrical conductivity at LJ01, LJ25 and LJ30.

At all three stations the conductivity decreased sharply in early May and continued to decrease until mid-May. The EC ranged from 20 to 40 μs/cm (LJ30 to LJ01) on May 12 to May 20th. After May 20th, the EC started to increase slowly at the lower and mid station, while at the upper station it remained constant.

Throughout June, the EC continued to increase at all three stations and each station had a very similar daily variation in EC. The diurnal signal of EC and discharge varied slightly in timing. EC at all three stations tended to peak a few hours before the daily low flow. Rain events in June did not cause a significant change in EC at any of the gauging stations, even though there were changes in streamflow. On June 28th, the streamflow at LJ25 was larger than at LJ01 but there is no indication in EC that there was any change in the composition of stream water at either station. The difference in EC from LJ30 to LJ25 remained constant at about 30 μs/cm and the difference between EC between LJ25 and LJ01 remained constant at about 20 μs/cm.

In July and August, the EC continued to increase at all three stations as streamflow decreased. Similar to June, there was a short lag between the peak EC and the lowest streamflow each day and there was a constant difference in EC between the stations, although the difference did increase by 10 μs/cm over the month at both LJ30 to LJ25 (40 μs/cm) and LJ25 to LJ01 (30 μs/cm). This lag between the peak EC to the daily minimum discharge became negligible towards the end of July at LJ01 and LJ30 and was reduced slightly at LJ25.

In August, the difference in EC between LJ30 and LJ25 increased to 60 μs/cm (70 μs/cm at LJ30 and 130 μs/cm at LJ25), while the difference between LJ25 and LJ01 increased by only 5 μs/cm to 35 μs/cm (LJ01: 165 μs/cm). The rain events during July and August caused a noticeable decrease in EC, unlike the rain events in June. Even very small rain events had a noticeable effect on EC (e.g., Jul 16 and Jul 19). This would suggest that streamflow was primary composed of groundwater and direct channel precipitation.

By mid-August there was almost no diurnal signal for the streamflow at LJ30. At LJ25 the diurnal variation in EC increased until August 21 when there was no longer a diurnal streamflow signal. Unfortunately, the conductivity probe was out of the water at LJ25 from August 25 to September 11th. At LJ01, the diurnal variation in EC was not as large as at LJ25 (although the streamflow diurnal signal at LJ01 was larger than at LJ25) and varied little in amplitude during August. Between August 20 and 28th, there were significant changes in daily streamflow at LJ01, almost reaching zero flow and yet there was no change in EC during this period. On August 26th flows became very low and EC measurements were affected by the lack of flow.

By mid September, the changes in EC transitioned at all stations from increasing to decreasing. The EC peaked at LJ01 on September 27th at 230 μs/cm, at LJ25 on Sept 11th at 180 μs/cm (approximate due to missing data), and at LJ30 at ~100 μs/cm in mid-September (missing data Sept 6th to 26th). The peak EC at LJ01 would suggest that there was some streamflow on the 12th of September but there was no measured flow for September 12th. This was also the case for the 18th, 27th and 28th of September. The peak EC occurred on a day that no flow was recorded, suggesting that instead these measurements reflect groundwater in the streambed. On Sept 11th the probe at LJ25 was put back in the flowing part of the stream and there was still a large diurnal signal. Field visits on Sept 12th and Sept 25th indicate there was pooling water at LJ30 but no flow. The September 26th rains resulted in a significant decrease in EC at LJ25, the only station with flow. On October 5th there was streamflow again at all three stations and but there was little to no diurnal variation in EC at any of the stations.

Over the month October, EC continued to decrease. This decrease was larger upstream (LJ01: 175 μs/cm to 165 μs/cm, LJ25: 125 μs/cm to 105 μs/cm and LJ30: 90 μs/cm to 55 μs/cm). The EC continued to decreased until the stream froze again. The diurnal EC signal was clear again at LJ01 and LJ25 in the latter half of October. On October 30th, the EC was about 10 μs/cm higher at every station than it was on March 25.

Manual EC measurements at other stream locations

EC measurements were taken at each streamflow measurement site along the stream (Figure 3.9). Overall they showed similar trends as the continuous measurements at the three gauging stations (LJ01, LJ25 and LJ30). In early May (3rd-5th), the EC between LJ30 to LJ00 ranged between 65 and 112 μs/cm, on May 9th between 34 and-94 μs/cm, and by May 12th between 31 and 61 μs/cm. The EC remained stable between May 12th and 20th (range: 35-77 μs/cm between LJ30 to LJ00), after which there was a slow increase in EC until the end of the month (range: 40-80 μs/cm). In May there was a gradual increase in EC with distance from the headwater, apart from sections LJ22 and LJ21 after the peak flow on May 27th, for which there was an increase in EC to 89 μs/cm and 79 μs/cm respectively. There were no increases in EC at LJ19, LJ13, or LJ07, which are located close to large hillside seeps with higher EC values.

On June 3, the range in EC between LJ30 to LJ00 was 42-81 μs/cm, with again LJ22 and LJ21 having a higher EC (81 μs/cm) than surrounding stations (67 μs/cm). This part was now almost a side channel and by June 6-7 the increase in EC was 30 μs/cm (to 85-100 μs/cm). This increase in EC at LJ21 and LJ22 is likely explained by the landslide (on May 25-26th) that caused the main streamflow to be diverted from these stations to LJ23 and LJ20 (which previously had very little streamflow).

Over the month of June and July there was a gradual increase in EC at all stations. By the end of June, EC values were similar to pre-melt EC (range: 54 to 110 μs/cm). By July 27th LJ22 and LJ21 EC values were similar to the main channel readings. There were no significant increases in EC at stations that were always gaining, suggestion that perhaps it is not deeper groundwater inputs that causes these sections to always gain but perhaps shallower streambed water that cause more water to be in the channel at these locations, or the groundwater inputs have the same EC as the other inputs.

In July and August, EC continued to increase: at the upper stations the EC was 140 μs/cm (LJ28) and at lower stations it was 235 μs/cm. The EC was similar in October, except that flow had returned to LJ30 and had a very similar EC to when it went dry (74 μs/cm).

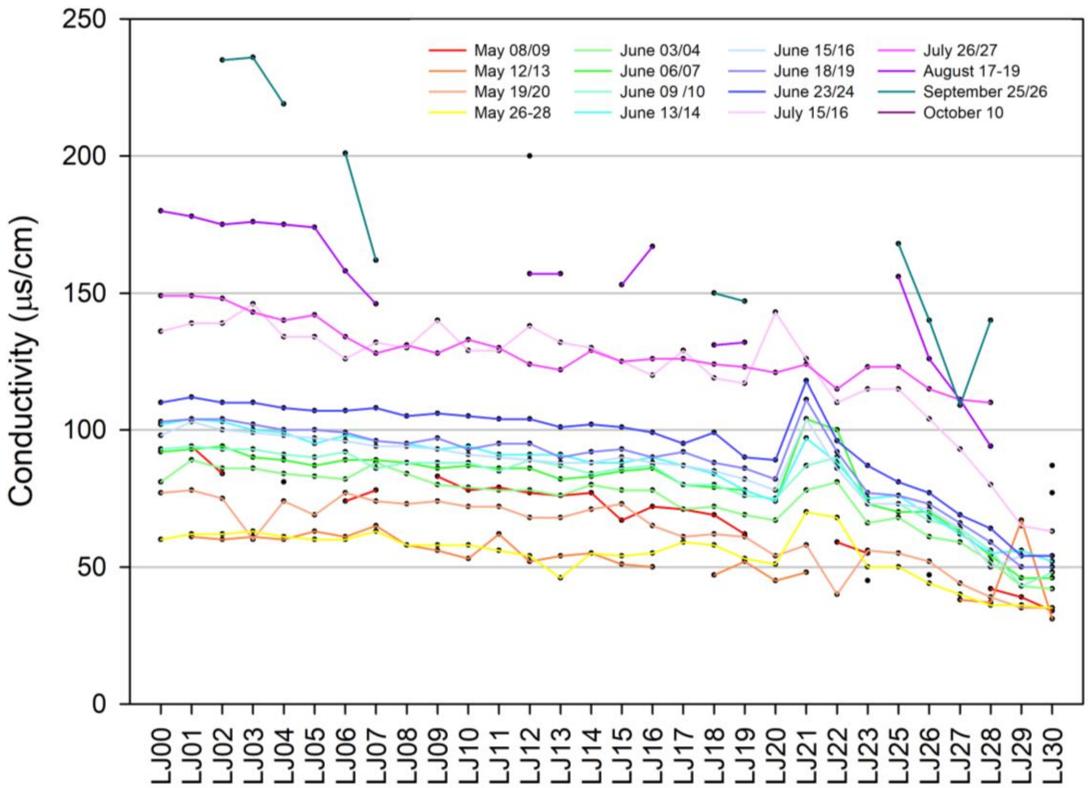

Figure 3.9 Manual conductivity measurements at streamflow locations during the field season.

EC in seeps and side channels

EC was also measured in the shallow groundwater, at seeps in the catchment and areas of extensive overland flow. The seeps in the lower part of the catchment near GWLJ02, GWLJ04 and GWLJ07 all had significantly higher (120 μs/cm to 150 μs/cm) EC than the stream EC in the nearby channel (100 μs/cm), suggesting that the residence time of water in these seeps is longer and that they could represent groundwater. The EC in these seeps followed a similar trend to stream EC with the lowest EC in mid-May. By early June, GWLJ04 had dried up and could no longer be used for measurements. The EC at GWLJ02 increased in early June (from 120 μs/cm on May 19th to 167 μs/cm on May 30th, to 180 μs/cm on June 3rd). The stream water EC also increased, but not by as much (May 19th:78 µs/cm, May 30th: 76 µs/cm and June 3rd: 89 µs/cm) . This would suggest that deeper groundwater was forced to the surface after the catchment became saturated. EC at GWLJ07 did not show this same large increase either; the increase in EC was comparable to the increase in stream EC (change from Mid-May to early June: GWLJ07: 20 µs/cm, stream: 10 µs/cm). In late June the effects of evaporation on the stream EC were noticeable: the EC of the stream water increased much faster than at GWLJ02 and GWLJ07. The EC of GWLJ07 (110 µs/cm) and the stream (108 µs/cm) were similar by June 23rd By July 17th,GWLJ07 was only a damp area. The EC at GWLJ02 had a similar trend over a longer period; the stream EC (221 µs/cm) was higher than GWLJ02 (208 µs/cm) by Sept 10th.

The seeps and side channels in the mid catchment followed the stream EC patterns closely. The side channels near LJ19 and LJ22 both had a similar EC as the stream throughout May until they went dry in early June. The EC of the hillside seep between LJ19 and LJ18 remained about 30 µs/cm higher than the stream until mid-July. The next reading at both stations occurred on September 25 and the stream (147 µs/cm) and GWLJ19 (143 µs/cm) had a very similar EC.

The seeps and channels between LJ30 and LJ25 had an EC (100-140 µs/cm) that was almost double that of stream water (60-70 µs/cm) in early May, which might explain why the EC increased so much between LJ30 and LJ25. The surface channels in the mid catchment (72 µs/cm) had a similar EC to the upper catchment (60 µs/cm), as did the seeps (upper and mid sites about 95 µs/cm, except for GWLJ02 which had an EC of 141µs/cm). Unlike at the lower locations, the EC tended to stay relatively stable between May 11th and May 15th when EC in streamwater decreased for most of the catchment. This suggests that the upper catchment seeps were less affected by surface water or that percolation is much slower in the upper catchment. The seeps were dry by mid-August and until then had and EC of 100 µs/cm (±15 µs/cm). The measured side channels in the upper catchment also maintained a relatively stable EC between 50-60 µs/cm until mid-July when they dried up. In early June, the upper catchment stream water had a similar EC as the side channels but unlike the side channels the EC continued to increase over July.

Groundwater EC

At LJ01 and LJ30 groundwater EC was continuously monitored at 50 cm below the streambed (Figure 3.10). At LJ01 stream and groundwater EC were similar until March 20th. From March 20 to 28th, the EC in the stream was 25 μs/cm higher than in the groundwater. On March 28th, there was a significant drop in stream EC; this drop occurred in the groundwater on April 1st. From April 1st to 30th, the groundwater had a lower EC than the stream. In late April, there was even a diurnal signal in the groundwater EC, suggesting it may be closely linked with surface water or the flow paths that are the source of the surface water. From early May until around June 21st, the groundwater EC at LJ01 was higher than the stream EC. There was little variation in the EC of the groundwater but there were diurnal variations in streamwater EC, suggesting that there was groundwater flow towards the bed. From June 21st until the end of July, the groundwater and stream at LJ01 had a similar EC and the groundwater EC also started to have diurnal variations but the signals were offset by approximately 11 hours. On August 21st, the groundwater piezometer went dry. This was 5 days before the stream started to go dry. On October 2, the groundwater piezometer again had water and the EC was similar to that of the stream but without a diurnal signal. These results suggest that the stream and groundwater at 50 cm below the stream were closely connected.

The piezometer at LJ30 started recording groundwater on May 1st. The groundwater had a significantly higher EC (45 μs/cm higher) than the stream and was almost the same as the EC in the stream and groundwater at LJ01. When the last snow melted (May 10th), there was a drop in the EC in groundwater (from 80 to 50 μs/cm). The daily average groundwater EC at LJ30 remained almost constant (although there were diurnal variations) until mid-June when there was a slow increase (to 63 μs/cm), which was also observed for the stream EC (to 50 μs/cm). The piezometer went dry by June 30th (until the records end on October 9).

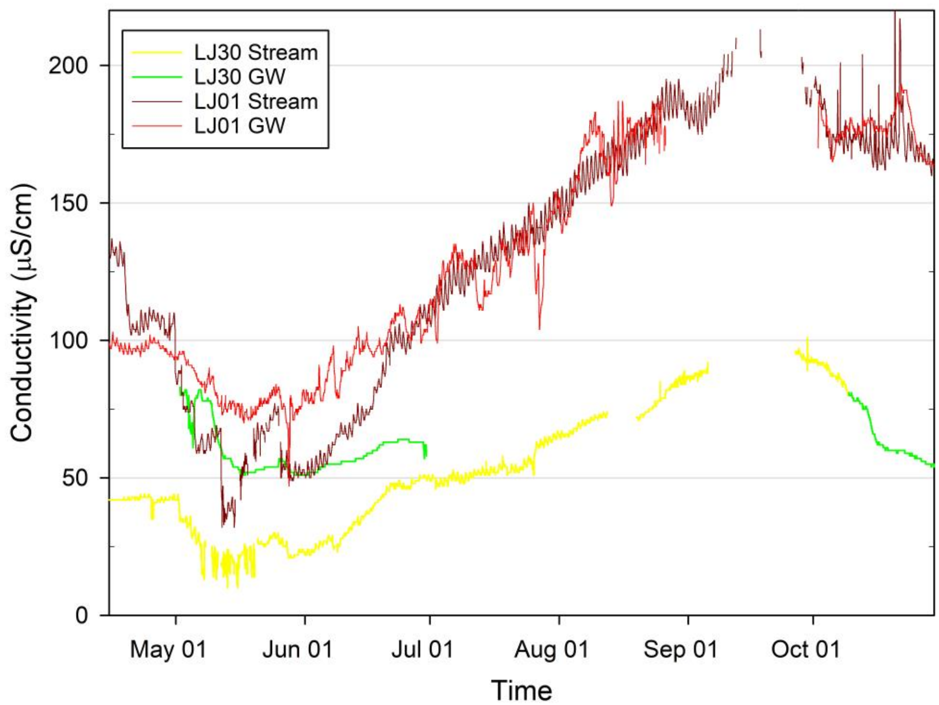

Figure 3.10 Electrical conductivity of streamflow and groundwater at LJ01 (red) and LJ30 (yellow). Large down spikes in late summer are due to the probe going dry.

3.5.5. Stream temperature data

Temperature probes were placed along the stream on the streambed (Figure 3.1) to determine the timing of streamflow (when flow started and stopped) and the spatial variation of streamflow (flow/no flow). The high failure rate of the sensors (only 20 out of 42 stream temperature sensors and 6 out of 9 air temperature sensors had data) was not anticipated and caused gaps in the results. Failures were due to destroyed or lost probes, buried probes, pools with no flowing water, flow around sensors and many probes failed to record data (probe defect). Some probes also failed after a short period of deployment. In the future more frequent downloads would be recommended to ensure the data are being recorded.

However, the stream temperature data recorded in March and April provided insight into the source of water at each location. In winter, warmer water likely comes from a greater depth, while colder water would have a source closer to the surface. Changes in the pattern of the diurnal variation of the stream temperature also provided an indication of the start of melt water flow. Large changes in the amplitude of the stream temperature signal would suggest the arrival of the melt water, while a small change in the magnitude of the diurnal signal suggested that there was a large groundwater component in the streamflow.

The spring melt water did not cause flow to immediately propagate downstream, rather there was large spatial variation in the initiation of flow (Figure 3.6;Table 3.2). Reaches that went dry first in the summer were not necessary the last to start flowing in the spring. The effects of groundwater in the stream caused stream temperatures to vary over the length of the catchment with no significant pattern in March and April, but by May there was a warming trend with distance downstream except for S5 and S16 which were clearly located at groundwater discharge locations. For a more detailed description of the spring temperature data, see Supplement A.

During the summer, the stream temperature trends started to diverge and no longer decreased with upstream distance. In July S19, S18, S14, S2, and S1 had large diurnal temperature variations and maximum temperatures were higher than at all other stations. Temperatures at S5 and S16 remained stable around 9°C and were just below the minimum temperatures of S20. S4 remained cooler than the surrounding probes and had a smaller diurnal signal. All other stations had very similar daily patterns.

By August, many locations in the stream started to go dry and the temperatures in the stream ranged from 9°C to the mid-20s. Determining exactly when the stream goes dry from the temperature data is difficult because the transition from flowing to dry was subtle in comparison to dry to wet during the melt period. It was also difficult to determine when flow started again in the fall (Table 3.2).

Table 3.2 Dates that flow started and or stopped at the locations of the temperature probes. If the diurnal air temperature and stream temperature variations had a similar amplitude and timing, it was assumed that the stream was dry. Flow=always flowed

Temp. Probe	Date Flow Started	Date Flow Stopped	Approximate location
S1	Flow	September 23rd (?)	Upstream LJ01
S2	Flow	Flow	LJ02
S3	Flow	Flow	Upstream LJ04
S4	Flow	Flow	Downstream LJ06
S5	Flow	Flow	LJ07 Left Channel
S6	March 28th	August 6th	LJ07
S7	April 1st	August 4th or 5th	Downstream of LJ09
S8	Flow	Pooling (July 31st)	Between LJ09 and LJ10
S9	March 13th	August 15th	Upstream LJ10
S10	Flow	September 10th - 12th	Between LJ10 and LJ11
S11	April 1st	August 3rd or 4th	Downstream LJ11
S12	March 24th	August 7th - 14th	Upstream LJ11
S13	March 27th*	Pooling (Sept 25th)**	Between LJ13 and LJ14
S14	March 28th – April 1st	August 3rd or 4th	Upstream LJ14
S15	March 28th – April 1st	August 9th or 10th	Between LJ14 and LJ15
S16	Flow	Pooling July 23rd - Aug. 2nd	Downstream LJ21
S17	Flow	-	LJ25 (until May 26th)
S18	-	July 28th – Aug. 1st	Downstream LJ23 (installed June 8th)
S19	-	Flow	LJ25 (installed June 8th)
S20	March 31st	August 4th	Downstream LJ29

*Pooling water prior to this date
**From field observations, could not tell from the temperature data

3.5.6. Comparison of air and stream temperature data and field observations

The diurnal variation in stream temperature could be used to determine the onset and cessation of streamflow at some locations but the signal was not very clear for other locations (Figure 3.11 and Figure 3.12). In some stream reaches, streamflow was so limited that it bypassed the sensor. In other locations, shading caused a relatively small variability in temperature despite the lack of water.

On April 1st, all stream temperature probes were in water and the standard deviation and the ratio of the standard deviation of the stream temperature and the air temperature were small (Figure 3.12 and Figure 3.13). The standard deviation of temperature for each day clearly illustrates the initiation of flow at stations with no flow (Figure 3.12), apart from S12 and S13. At S12 flow started on March 24th and S13 was pooling prior to March 27th. The ratio of the standard deviation of stream temperature and air temperature also illustrates the initiation of flow at all stations with no flow except for S20; S5 had flow but appears dry until April 1st. This may be due to pooling water. The low ratios in early March for stations with no flow, likely indicates the presence of snow/ice and the timing of the melt can be estimated based on the increase in the ratio.

During the summer, stream temperature variations were more similar to air temperature variations for many probes, indicating that the streambed could be dry (Figure 3.18). Exactly when the stream went dry was difficult to determine from the standard deviation or the ratio of standard deviation, in part due to the nature of point observations. Probes that consistently had very low ratios of the standard deviation were likely influenced by groundwater inputs that stabilised the stream temperatures, for example S5, S13 and S16.

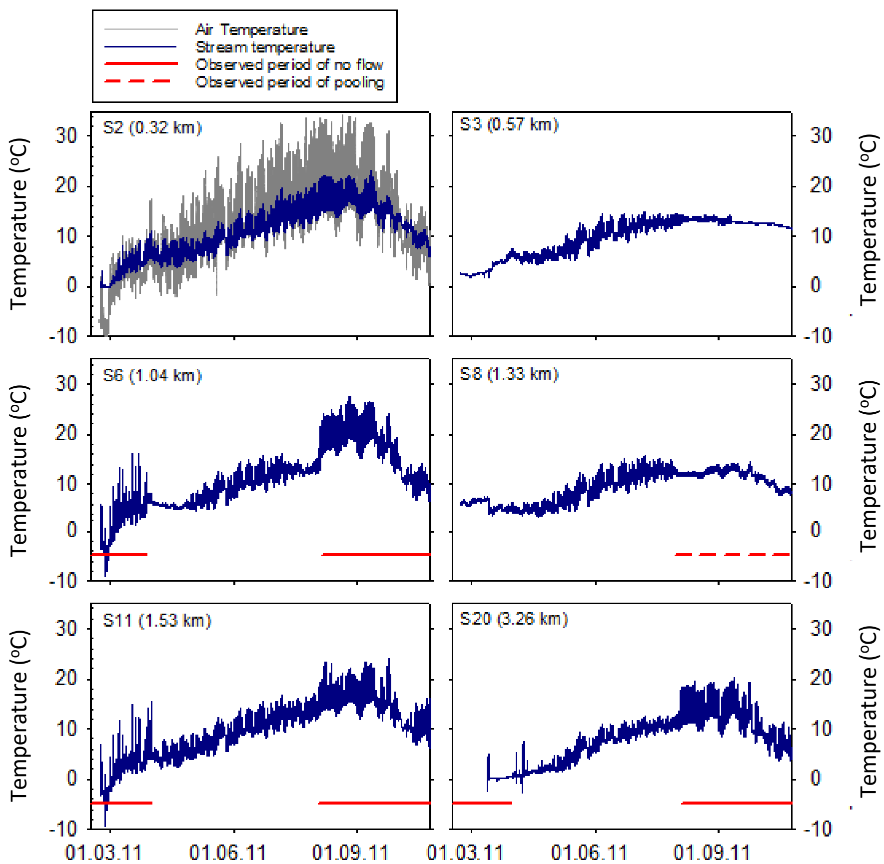

Figure 3.11 **Stream temperature and air temperature variations, compared to observations when the stream was dry (red line) or pooling (red dotted line).**

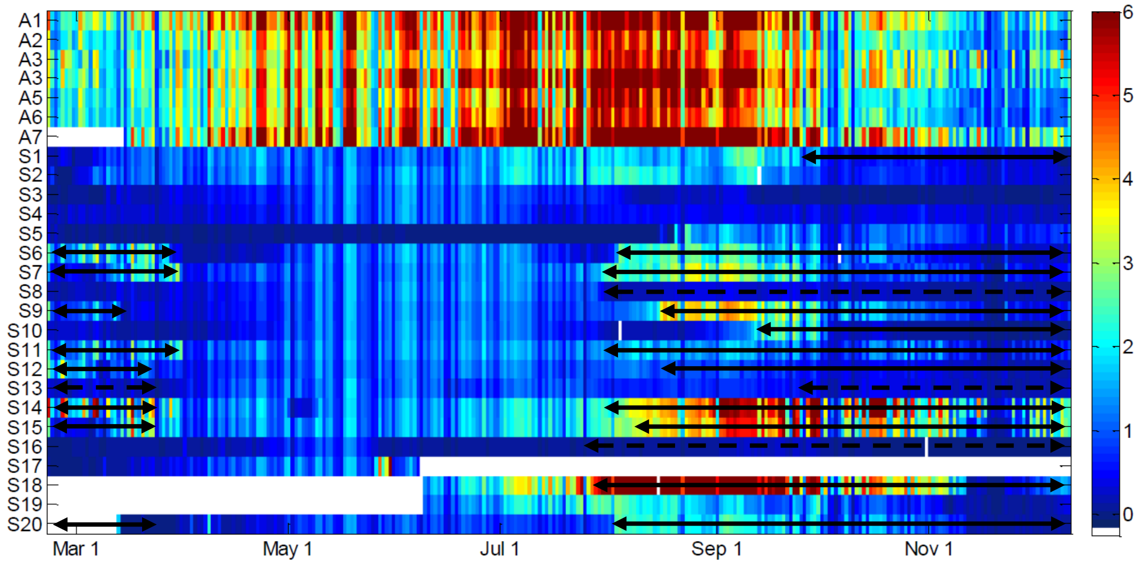

Figure 3.12 **The standard deviation of stream temperature and air temperature for each day and observations of when the stream was dry (black arrow) or pooling (dashed arrow).**

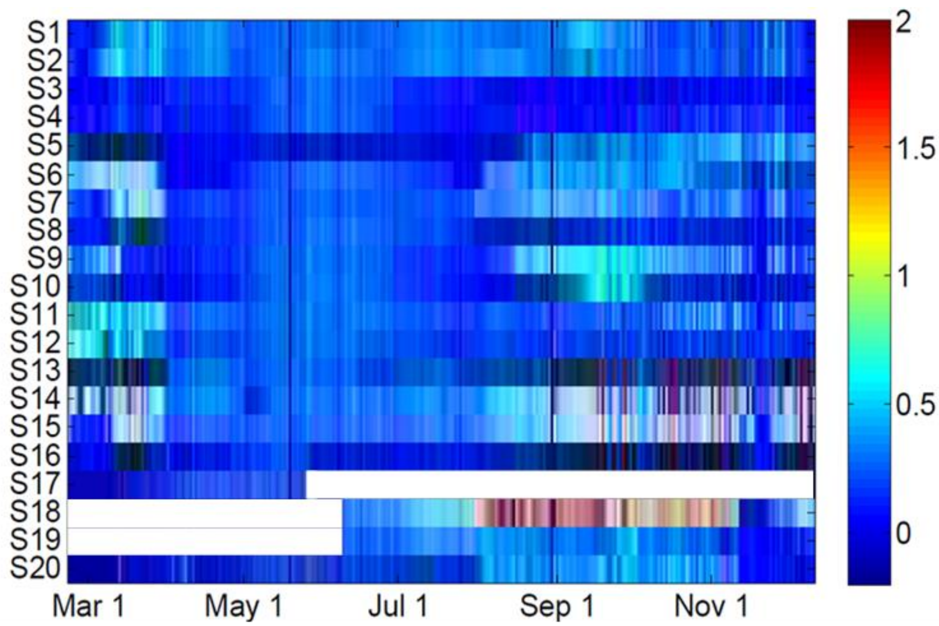

Figure 3.13 **Ratio of the standard deviation of stream temperature and air temperature for each day**

3.5.7. Temperature profiles

Subsurface temperature profiles

At three locations temperature was measured every 5 minutes over a range of depths (LJ01, LJ22 upstream and downstream). These measurements provided an indication of whether the stream was gaining or losing. The propagation of a large diurnal signal to depth suggests a losing section, while the lack of a diurnal signal at depth would suggest a gaining section. Diurnal temperature variations also provided an indication of presence or absence of water at depth.

The data from the four probes at LJ01 suggest that this was always a losing stream section because he deepest probe always had a diurnal signal (Figure 3.14). In early March and after mid-September, the temperature profile was inverted and temperatures were highest at the deepest sensors, as would be expected because the air temperatures were low. Water level data from the stream and piezometer confirm that this is a losing reach for the period of data collection.

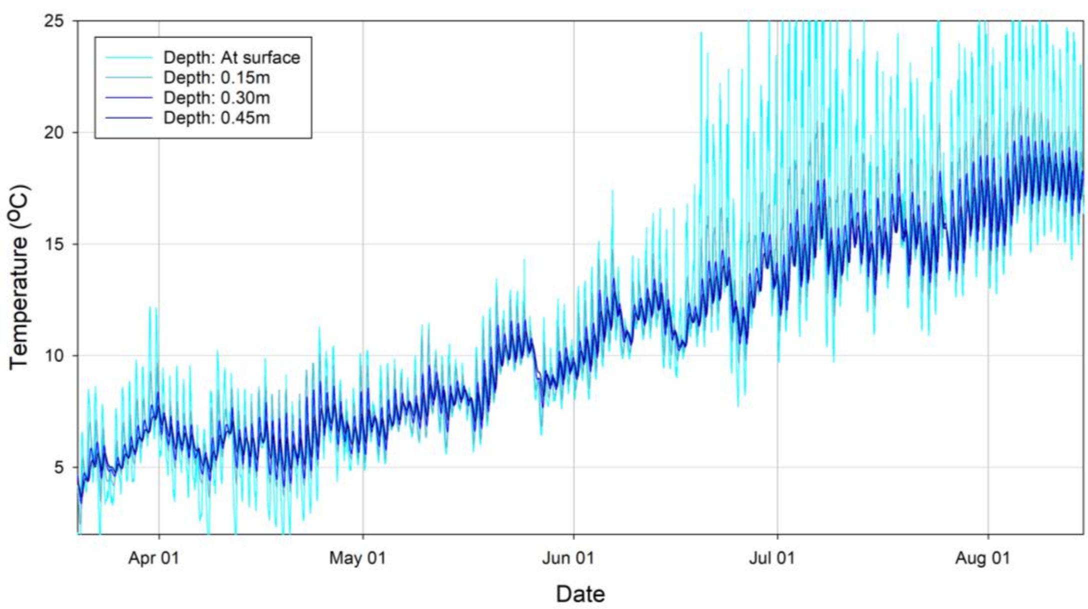

Figure 3.14 Temperature profile data for LJ01. The site was always a losing reach.

The data from the LJ22 upstream profile suggest that this area was a gaining stream until April 1st (Figure 3.15). There was almost no diurnal signal at 0.38 m until April 1st.

After April 1st the diurnal temperature range probe varied between 0.2°C (April) to 0.7°C (late August). This very small variation in daily temperatures in comparison to the diurnal temperature fluctuations at 0.01 m (1°C in April to 8°C in August) suggests that this was likely always a gaining stream section. Piezometer and stream water levels indicate that it was a losing reach between May 9th to June 7th but this is not easily discernible from the temperature data. In February this area was snow free and very damp; it remained damp on the surface throughout the summer.

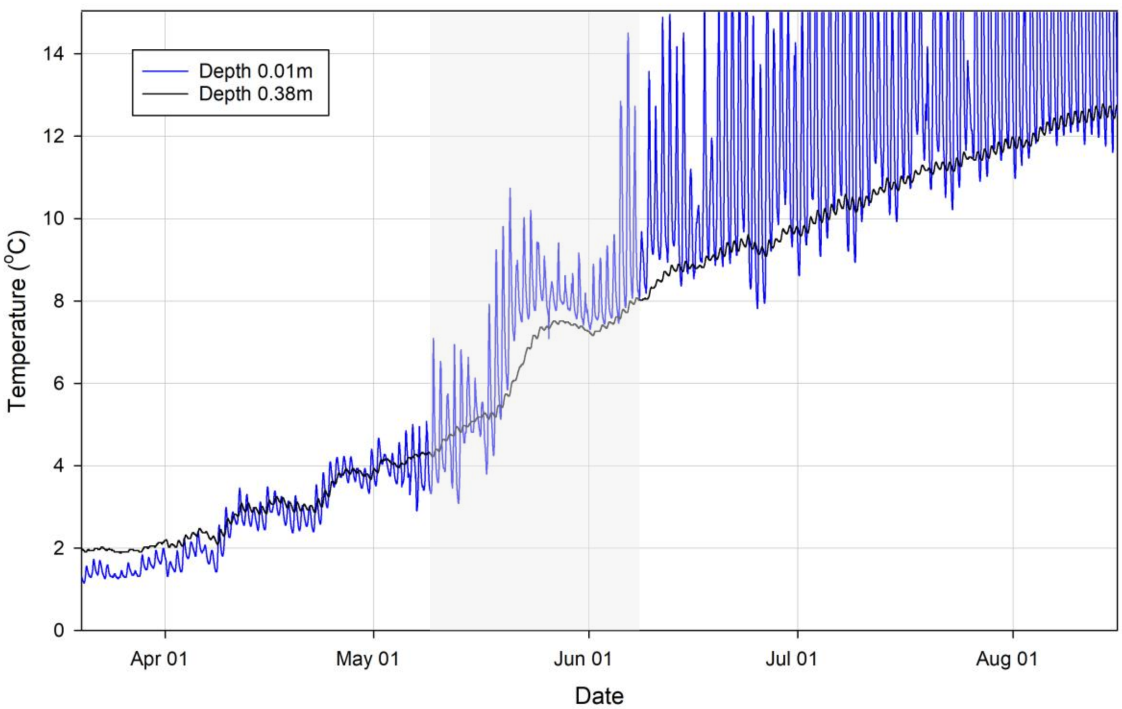

Figure 3.15 Temperature profile data for the location upstream of LJ22. Grey area represents the period when the reach was losing according to the water level data.

The upper part of the lower mid profile downstream of LJ22 was frozen until April 1st (Figure 3.16). After April 1st there were large diurnal variations at all depths, indicating that it was a loosing reach. From mid-May to late June, the very minimal daily variations in temperature suggest that the reach could be gaining flow but the piezometer and stream water levels suggest that the reach was still losing. On August 15th the upper probes went dry; the deepest probe followed 4-5 days later.

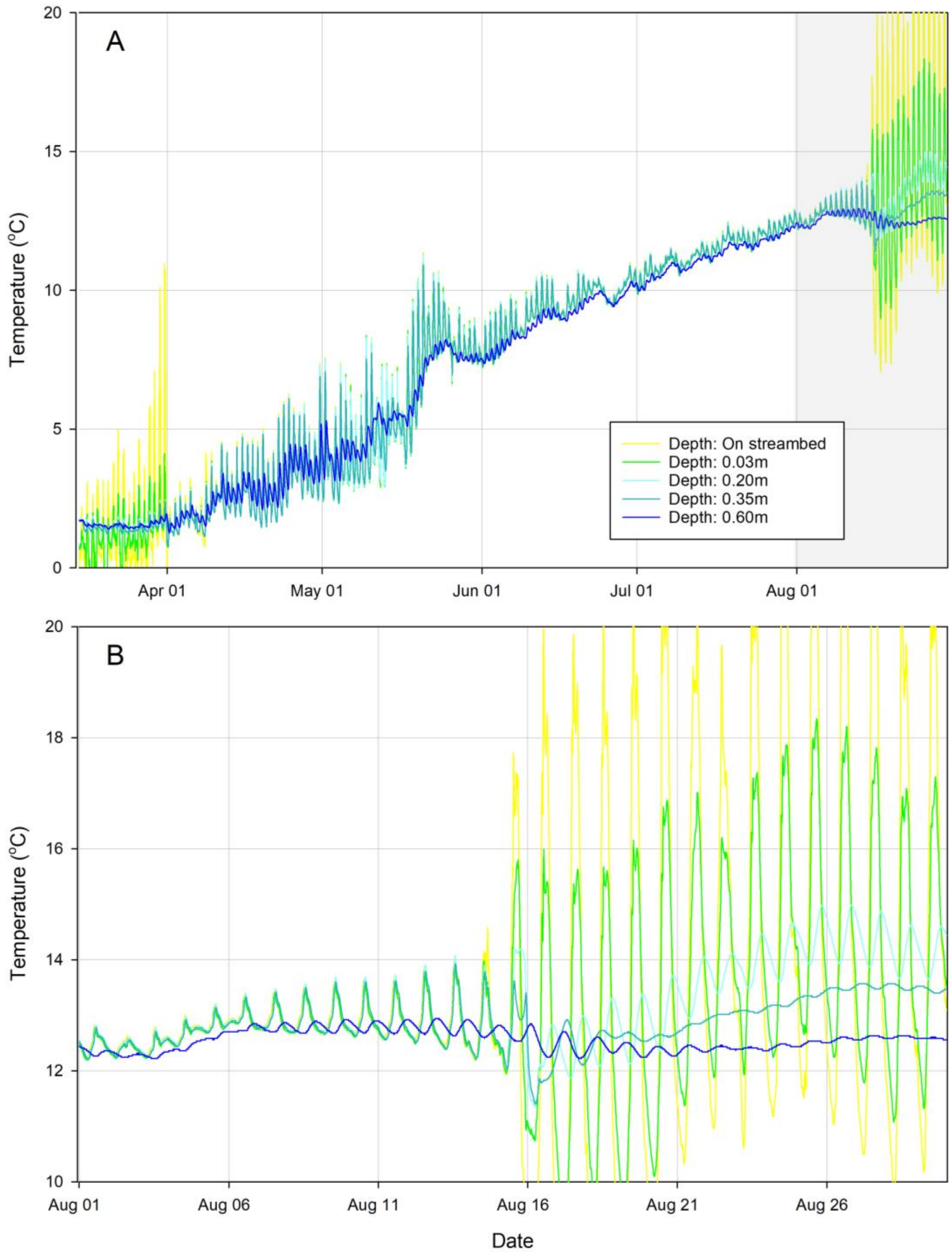

Figure 3.16 Temperature data for the profile downstream of LJ22. B focuses on the August data (shaded area in A).

3.6. Discussion

3.6.1. Gaining and losing stream reaches along Long Joe Creek

The findings of this study highlight that the contributing catchment area is not a determinant for discharge. In May and June, the increases in discharge in the upper portion of the catchment (LJ30 to LJ25) were much greater (50-200% gains) than for the lower portion (LJ19 to LJ01) of the catchment (2-20% gains). Downstream of LJ25, post freshet but pre-dry out, there was an increase in streamflow until LJ20 and then a stabilization of streamflow, even though there was a 40% increase in drainage area. This suggests that the relative contribution of the lower part of the catchment to streamflow was small or largely offset by transmission losses. Bergstrom et al. (2016) and Payn et al. (2012) also found that increases in contributing area did not necessarily relate to increases in discharge, especially, as the watershed dried the contributing area had less of an effect on discharge.

The differential gauging results indicate numerous regions that were losing but it is difficult to determine if the losses were to groundwater or streambed flow that resurfaced further downstream. The overall basin trend of losses below the winter fog line suggest that there were large permanent losses to groundwater from the channel. These results are similar to other studies which found larger losses to groundwater downstream, especially during transition periods from wet to dry and dry to wet conditions (Bernal and Sabater 2012; Butturini et al. 2003). However, the results are contrary to the findings of Wassenaar, Athanasopoulos, & Hendry, (2011), who concluded based on isotope analyses that streams on Anarchist Mountain were not a major source of recharge for bedrock.

A study using differential gauging for small perennial streams (using velocity-area flow calculations) suggests that large scale measurements provide a better estimation of groundwater recharge than point measurements because the error in flow measurements over short distances was larger than the difference in flow (Cey et al., 1998). These results from Long Joe Creek, suggest that there is a large variability in the gains and losses to groundwater and that point measurements are very useful and provide a detailed understanding of where the stream loses most water and what type of events cause the greatest losses or gains in streamflow. If we did not have data for

LJ25, the catchment could have been assumed to be a losing catchment for much of the freshet, yet our data shows there are significant gains above the winter fog line that are lost over a very short distance below the winter fog line.

The winter valley fog line (located 100 m above the fault line) has a large effect on streamflow and water availability in Long Joe Creek. Gaining reaches in Long Joe Creek were primarily located above the valley fog line (LJ19 and above) in an area of greater snowfall and snow persistence, but below the flat upper headwater area. Overland flow was extensive during the freshet and post-freshet in the upper reaches of Long Joe Creek and these areas continued to have saturated streambanks and flowing channels after snow melt. Flow increased consistently more than 5% between each station between LJ29 to LJ26 over the course of the year but the volumes of the gains decreased as the summer progressed the catchment dried out and temperatures increased. This suggests that the source is primarily melt water that was being depleted due to the continued drainage. The sharp increase in conductivity over this reach also suggests that groundwater is the primary water source. The drying of LJ30 and LJ29 suggest the groundwater levels were significantly lower during the summer months than during freshet and were no longer a source of stream water in the upper basin. Alternatively, the decrease in the gains can be due to increased ET from the dense stand of coniferous trees and shrubs in the upper part of the catchment. A study by Bergstrom et al. (2016) in the Tenderfoot Creek Experimental forest, Montana, USA also found that streamflow in the upper basin of numerous watershed channels contracted during the summer months in the upper reaches.

The section between LJ25 and LJ28 always had flow, even when LJ29 and LJ30 went dry. The consistent streamflow increases between LJ28 and LJ25 indicate that there was a larger storage capacity in the soils in this part of the catchment, the groundwater table was close to the surface and that water from the upper portions of the catchment accumulated in this area. During freshet and early post-freshet, this reach had saturated soil conditions and overland flow was observed frequently.

The increase in streamflow between LJ20 and LJ19 could be due to groundwater discharge between these stations. This section was flowing during the winter and damp throughout much of the summer. This groundwater discharge could be related to the fracture and transition from granodiorite to Anarchist Schist bedrock (Figure 3.1) near this location. Fractured rock and lithologic contact zones can create areas of increased permeability and therefore presence of groundwater (Crawford and Brackett 1995) or block further groundwater flow and cause exfiltration (Huff et al. 1982).

There were also a few small gains in the lower sections (LJ13, LJ07, LJ02 and LJ03) but overall the stream loses water across this lower and drier part of the catchment. Determining if these losses are to the regional groundwater or subsurface flow is difficult. In cases where the losses and gains occurred over a short distance the most obvious explanation is flow through unconsolidated streambed material that has sluffed from the surrounding banks (and partly old landslide deposits in the stream). The reaches between LJ14 to LJ15 and LJ08 to LJ07 are very good examples of reaches that had persistent losses and gains due to flows in unconsolidated material below the bed. LJ14 and LJ08 consistently had losses of 5% or more and LJ15 and LJ07 had gains of 5% or more and the measured volume of gains and losses were almost equal to each other and remained constant over the season (Figure 3.4). If the differences were due to measurement errors, the patterns wouldn't have been so constant from one measurement campaign to the next. But there must also be a groundwater gains, or inter flow from higher in the catchment, because LJ08 went dry while there was still flow at LJ07.

There appear to be permanent losses to bedrock or the streambed between LJ11 and LJ08 as gains in flow were never observed. This region was also the first to go dry during the summer, which indicates large streambed losses. The bedrock was primarily Anarchist Schist which can have a higher porosity. However, LJ07 always had the largest gains in the lower portion of the catchment. Thus, it could also be that there is significant flow through the channel bed between LJ11 and LJ08. Flow at LJ07 was often similar to the flow at LJ12.

The streamflow losses at LJ05 were not expected as flow was always present at LJ04 and LJ06 and between LJ06 and LJ05 a horsetail patch persisted over the season, suggesting groundwater exfiltration. Even with this input, LJ05 went dry and nearly always had lower flows than LJ06.

LJ01 is located near what would have been the apex of the alluvial fan if urban development had not taken place. The channel between LJ03 and LJ02 is bedrock

(Alkali feldspar granite). At the transition from bedrock to alluvial fan losses to deeper unconsolidated material could explain the consisted streamflow losses post early freshet between LJ02, LJ01 and LJ00. However, the lower and upper catchment have a similar mean D50 (0.108 m and 0.107 m respectively) suggesting that the larger losses in the lower catchment were not solely due to differences in porosity, although the bed material is very heterogeneous.

The streamflow at LJ00 was the largest in the catchment at peak flows but at all other times maximum streamflow occurred upstream, often at either LJ02, LJ03 or LJ12 until very late in the season when peak stream flows occurred further up in the catchment. The very small gains in the lower basin during freshet are more likely due to smaller contribution from groundwater. They could also be due to the higher ET on the lower slopes and less snowmelt recharge that kept the soils relatively dry and limited subsurface flow contributions to the stream. In the lower catchment, there were also large transmission losses, even at high flows. These findings could explain why the peak discharge migrated upstream as the summer progressed because less streamflow meant relatively higher losses to groundwater in the lower part of the catchment. This groundwater recharge suggests that small tributary drainages should not be ignored during water balance calculations of larger catchments in arid and semi-arid regions and that streamflow should not only be measured at the outlet of the catchment. Using differential gauging in an arid environment, Mccallum et al. (2014) revealed that smaller events were more important than large events for groundwater recharge; even though the rates of recharge were low and the flows were small, the total volume of recharge was greater than for larger flows because these flows lasted longer.

3.6.2. Patterns of flow and no flow

Above the fog and fault line, the regions of the streambed that were dry in the winter were also the first to dry in the summer. Below the fog line, regions that were dry in the winter did not necessarily go dry first. All regions that were dry in the winter did eventually go dry in the summer. During the winter, snow accumulation above the fog line (~1020m) was significantly higher than below the fog line, where snow often disappeared between events. This larger accumulation of snow above the fog line contributed to spring streamflow gains in the upper reaches and the slow release of this melt water may have contributed to the persistent gains in the upper regions later in the spring and summer.

As the catchment dried, there were some obvious patterns in flow that developed. There was simultaneous drying from the outlet in the upstream direction and from the headwaters downstream. Areas that tended to lose water went dry first and these dry areas expanded, but never coalesced completely. This led to a high variability in the occurrence of water along Long Joe Creek, with up to 35 transitions between flow, no flow and pooling in late summer (Figure 3.6).

For example, above and below LJ12 and LJ13 there were extensive dry areas, but LJ12 and LJ13 remained flowing. This could be explained by a groundwater input at LJ12 and LJ13 that was almost equal the to ET and streambed losses and therefore did not flow far downstream. Another explanation could be that large volumes of porous bed material between LJ17 to LJ14 caused the flow to occur primarily through the streambed and to then resurface at LJ12 and return subsurface downstream further down. Payn et al. (2012) concluded that upstream transmission losses can travel along subsurface flow paths and resurface downstream and increase the expected streamflow at a given location.

The bed material appears to be the only stream characteristic that is useful to predict whether a reach will go dry or not. The six reaches with the largest D50 of the bed material (LJ09, LJ11, LJ13, LJ20, LJ21, LJ00) and the largest D84 (LJ08, LJ11, LJ13, LJ14, LJ21) all went dry. Thus, if the D50 was greater than 0.168 m and or the D84 was greater than 0.620 m the streambed went dry. However, fine bed material (small D50 or D84) does not provide an indication of whether a reach will have flow or not, as many reaches with fine bed material went dry as well. Slope or bankfull flow were not indicative of flowing or drying stream sections. However, Levick et al. (2008) states that intermittent and ephemeral channels rarely reach “process-form equilibrium” and bedforms rarely represent flows. Therefore, if calculations are done based on measurements of these bedforms misleading results will be found.

Ten stations had continuous streamflow over the entire duration of the study (LJ03, LJ04, LJ06, LJ07, LJ12, LJ18, and LJ25-LJ28). LJ08 was one of the first stations to go. In the reach between LJ08 and LJ07 water could be heard beneath the surface and there was a groundwater seep to the left of the channel. This suggests that the flows at LJ07 were consistent due to groundwater inflows and or subsurface channel(s) as described by Payn et al. (2012). At LJ18 the channel banks were steep and narrow; the streambed was on bedrock. It can only be assumed that there were little to no losses to bedrock and therefore flows were continuous. The continuous streamflow at the other sites are not so easily explained but were likely also do to exfiltrating groundwater.

There were nine reaches that had flow in the winter but no flow by the end of summer. LJ01, LJ02, LJ15, LJ16, LJ19, LJ30, S16, a small area between LJ15 and LJ16 and between LJ28 and LJ29. LJ01 and LJ02 may have had no flow at the end of summer due to high evaporation. At LJ15 and LJ16 and the region in between a large amount of new unconsolidated material was deposited during the high flows at the end of May and this likely caused larger losses to and flow through the streambed in this reach. This was also likely true for LJ19. S16 was installed at the bottom of a large drop below LJ21 in an area of warm water. This was assumed to be a groundwater discharge area, and therefore the groundwater table must have dropped or the source water for this area was no longer available. The area between LJ28 and LJ29 and the reach at LJ30 all went dry but there is no obvious reasons for them drying in summer and having flow in the winter. Perhaps the drying could be explained by increased evapotranspiration in the hotter drier late summer.

3.6.3. Usefulness of temperature measurements for determining the onset and cessation of flow

Previous studies have shown that temperature can be used to determine the timing and duration of intermittent or ephemeral streamflow (e.g. Constantz et al. (1996), Constantz et al. (2002) and Blasch et al. (2004)). These studies all focused on temperature at selected locations and then extrapolated the results. The disadvantage of temperature profiles is their inability to capture the multidimensional spatially variable nature of streamflow losses beneath temporary streams. The advantages of temperature profiles are their robust nature and the continuous data on the patterns of intermittent flow.

A large difficulty for the temperature profile measurements was finding soils deep enough to install the sensors. Previous research (Constantz et al. 1994, Anderson 2005, Constantz et al. 2003) using temperature profiles mainly focused on homogenous sand aquifers so that installation was not an issue and allowed for quantification of flow (Niswonger et al. 2003, Bravo et al. 2002, Constantz et al. 2003). Quantification of water fluxes from temperature data in heterogenous substrates, such as in Long Joe Creek results in very large uncertainties in the calculated fluxes. The three profiles, we were able to install provided both an indication of the presence of water and an indication of the direction of flow.

The streambed temperature measurements were used to determine the timing and spatial variation of flow and were compared to the observations. The streambed sensors were easier to install than the profiles but the loss rate was high. The temperature data provided a good indication of the source water. Warm water in winter and or a minimal diurnal signal during summer were good indicators of groundwater. Initiation of flow was also easily determined from the temperature measurements due to the large change in the signal. Temperature was also very useful to determine periods of no flow, when the signal matched of exceeded air temperature. A main issue to determine the cessation of flow was related to whether the probe was installed in the lowest part of the streambed or not. This is difficult to guarantee in temporary streams with mobile bed material and high sediment transport rates. The exact timing might be off due to flow not covering the probe or the probe being buried by sediment. Also, field observations and water level measurements indicated that the stream would change states numerous times from flow to no flow prior to going dry completely. Teasing these variations out of the temperature signal was difficult. It is therefore useful to combine stream temperature measurements with water level and conductivity measurements or streamflow measurements and the mapping of the occurrence of flow (as done in this study) to determine the occurrence of intermittent streamflow.

3.7. Limitations of the study

There are numerous limitations to this study. These may have led to some of the discrepancies between the results from the different types of measurements (e.g. the water level data suggesting losing conditions while the temperature profile data suggested gaining conditions at the upper-mid station). This highlights the importance of combining different types of measurements.

Streamflow data at LJ01, LJ25 and LJ30 have an inherent uncertainty due to uncertainty in the stage-discharge relationships. These errors tend to be particularly large when gauging a small creek with mobile stream bed material. Water level sensor drift was corrected for by distributing the correction over the period of record as no instantaneous corrections were obvious, and streamflow measurements plotted within 5% or 1 mm of the stage-dicharge curve. The volume of flow being measured was very small and therefore small measurement errors can have a large effect on the results. Some of the RWT for the point measurements may have been absorbed or stored in the streambed or the bank. Gains and losses also occur simultaneously and may partially cancel each other, therefore the gross exchanges are not measured (Covino and McGlynn 2007; Payn et al. 2009).

As discussed previously there were numerous issues with the temperature probes, including being buried, sitting in a pool of water or being dry because the water flowed in a different location of the streambed.

Finally, this study was done over one field season in one catchment, with a particularly wet freshet period. Additional monitoring would provide more insight on the typical duration of the dry streambed, and perhaps also the influence of topographic and catchment characteristics on the occurrence of flow.

3.8. Conclusions

The occurrence of streamflow along Long Joe Creek was highly variable and depended partly on topographic and catchment characteristics. The fog and fault line located near LJ21-LJ23 defined the overall transition from gaining to losing stream sections; below this point the stream was largely a losing stream and there were only small gains in streamflow. The catchment dried in a fragmented pattern. The first regions to dry were located in the middle and upper regions of the catchment, followed by drying at the bottom and top and an expansion of the middle dry reaches. The expansion involved some coalescence of dry reaches to create large dry reaches but in other areas short reaches went dry with little to no expansion.

Flow through unconsolidated material in the streambed from surrounding hillslopes could cause reaches to go dry due to losses to the streambed. All sites with coarse streambed material (large D50 and D84 values) went dry. The slope of a reach did not have a pronounced effect on the presence or absence of flow. Groundwater contributions from seeps maintained flow in summer in some regions. But in other locations where groundwater flow kept the stream flowing in winter, the stream went dry in summer.

Temperature data were useful to determine whether there was flow or no flow in the stream but there was a very high failure rate of the probes and therefore loss of data. The probes that did work provided accurate information on the timing for the start of flows but less detailed information on the transition to dry periods. The exact timing that the stream went dry was difficult to determine from the temperature data, partially because the stream did not go dry instantaneously but changed state numerous times before drying completely. There were also probes that were buried in the streambed material, located in ponded water, or where water flowed along the probe. The temperature probe data thus needs to be compared with field observations to provide reliable insight into the timing and spatial distribution of flow.

Acknowledgements

I would like to express my deepest appreciation to all those who helped me to complete this study and acknowledge with much appreciation the fieldwork support I received from Emily Garlich, Liz and Terry Huxter, Andrea Vowell, Jennifer Wilson, Shayne Doerksen, Sheena Spencer. Special thanks to the invaluable help Natasha Murphy provided in the field. Her positive attitude, work ethic, quick thinking and attention to detail made fieldwork both enjoyable and worthwhile. Without access to Jim Pentagraphs' land this study would not have been possible. I would also like to thank Environment and Climate Change Canada for the equipment used to carry out this study.

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3.10. Supplement A

Detailed description of the spring temperature data

Before April 1st stream temperature did not decrease with distance upstream due to groundwater being the main source of streamflow. However, in early April most upstream stations tend to be cooler than the downstream stations, likely due to the presence of melt water flow at all stations. In April streamflow started at all stations that had no flow earlier in the winter (S6, S7, S9, S11-S15, S20).

Stream temperature changes were often not uniform or predictable. During March and April some stations had very different patterns from nearby probes yet similar patterns to distant probes, there were also probes that had similar trends until melt water arrived and then had very different responses to the melt water. S1 located 23 meters upstream of LJ01 and S2 located at LJ02, both had flow during the winter, and very similar temperature patterns over the entire season. In early March stream temperatures at S1 and S2 were considerably lower than most stations, other than S17. Although starting April 1st, the daily maximums (9°C) at S1 and S2 were the highest in the catchment. The daily minimum temperatures were similar to the peaks of probes upstream of S7. S3 (just upstream of LJ04), S4 (just downstream of LJ05) and S13 (between LJ13 and LJ14) record a stream temperature of 2°C in early March. The temperature at S3 and S4 increase during the month; the temperature at S13 also increases but at a much slower rate (2°C cooler than S3 or S4 April 1st). S5 (left channel upstream of LJ07) was very warm at 7.3°C likely due to groundwater inputs. During the month of March, all stations downstream of S5 are warming, but S5 had a cooling trend. S5 and S16 had the highest winter temperatures (7.4°C) and almost no daily variation in temperature, this also suggests the presence of groundwater inputs. On March 14th and 15th, the temperature at S5 to 7°C decreased (one day after the large temperature drop at S8) and then continued a steady decline with little to no daily variation. Unlike S5, S16 has a significant one day drop of 4°C on April 1st. Post drop the temperature at both sites decreased until April 25th (down to 5.6°C) when there was a 0.25°C increase in temperature. On March 29th probes S1 to S6 and S16 had similar temperature signals. The amplitude decreased upstream, but all had similar minimum temperatures (~6°C), mid catchment minimum temperatures were about 2°C lower. S6 started flowing March 28th which caused a large rise in temperature (4.5°C to 6°C just after midnight) and a decrease in the diurnal variation (from >3°C to 0.25°C). S6 had a different trend than upstream stations until April 8th: it was about 2°C warmer than the mean temperature of upstream probes, was always warmer than the peak temperature of the upstream stations, the diurnal temperature variation was only 0.25°C each day and temperatures decreased with time (the cooling trend was similar to S3 and S4). After April 8th the daily variation increased to 0.5°C and the temperature was close to the mean temperature of upstream stations which varied by 3°C each day. When S6 started flowing it had the same temperature pattern as S5 until April 12th and afterwards was cooler by 0.3°C and had a larger daily variation (0.5°C) (similar to S4). S7 appeared to start flowing April 1st based on a large reduction in the diurnal temperature variation. The daily temperature patter and S7 is similar to S8 to S10. S8 had flowing water between 6°C and 7°C until the night of March 13th when the temperatures dropped to 4.5°C -5°C over three hours and then followed the S7 and S9 to S10 temperature trends but remained a degree warmer. S9 was assumed to start flowing at 21:00 on March 13th based on a large rise in temperature during the night and then had temperature variations that followed S10, although they were slightly larger (0.3°C). S10 (4°C) and S13 (3°C) had very similar daily variations (~1°C) until March 27th when S13 started to warm and vary by 2°C or more daily, while S10 remained stable with a 1°C daily variation. S12 had no flow until March 27th and then followed the same daily trends as S13. At S11 flow started on April 1st and the temperature was similar to S13 and S12 from then on. S13 was installed in a pool and likely did not have flow until March 27th. S14 and S15 both appear to have started flowing between March 28th and April 1st and had a similar daily pattern. As stated before S16 and S5 had very similar temperature variations until April 1 when the temperature at S16 dropped 1°C and became more similar to that at S10. Over this period the minimum temperatures increased and maximum temperatures decreased, similar to S12. S16 appeared to be warmer than most probes downstream, with a minimum temperature of 4°C. S17 was installed in flowing water with temperatures close to 0°C (like S1 and S2); the temperature continued to increase during March and by March 15th a diurnal pattern was visible as well. Probes downstream of S17 measured a stream temperature of 3°C or warmer during March and 1.5°C to 2°C warmer during April but have a similar daily pattern. The daily minimum temperatures at S17 were very close to 1°C throughout April. The maximum temperatures increased but the minimums stayed relatively stable throughout the melt period. S20 was located near LJ29 in the upper catchment in an area of no flow. The temperature remained stable at 0°C until March 31, when a diurnal signal started and was considerably muted in comparison to the air temperature and was assumed reflect the appearance of flowing water, although on April 3,7,8,11 there were large spikes in temperature suggesting there might not have been continuous flow or flows were very small. Temperatures at this probe were the lowest in the catchment remaining near 1°C until April 15th. After April 15th the minimum temperatures at S20 were similar to those at S17 (located at LJ25). There was a daily temperature range of about 1°C even at the end of April, which was considerably less than most downstream stations for which it was 2 -5°C.

Between May 1st-25th, stream temperatures were more similar and more predictable. There was a cooling of stream temperature with distance upstream and a diurnal signal range of about 5°C at all stations, except S5 which had a very stable temperature of 6°C that increased slowly over the month. After the large rain event May 25, sensor S17 was removed due to debris cover. S16 was installed in a pool of warm water in the winter, after the large event (May 25th) nearly all the flow was in the side channel (LJ23 and LJ22) not the LJ20 and LJ21 channel. The temperatures in the pool at S16 were very stable at about 7.8°C which was about 0.5°C warmer than S5, which was another pool of warm water in winter.

In June, there were smaller differences in stream temperatures across the catchment. On June 1st S5 and S16 (8°C) had a similar temperature as the minimum temperature at S6 to S19 and didn't change, while all other stations except S20 continued to warm. In early June, the maximum temperature at S20 was similar to S5 and S16 (~8°C) but S20 rapidly started to warm and by the end of the month S5 and S16 had a temperature that were more similar to the minimum temperature at S20 (8°C).

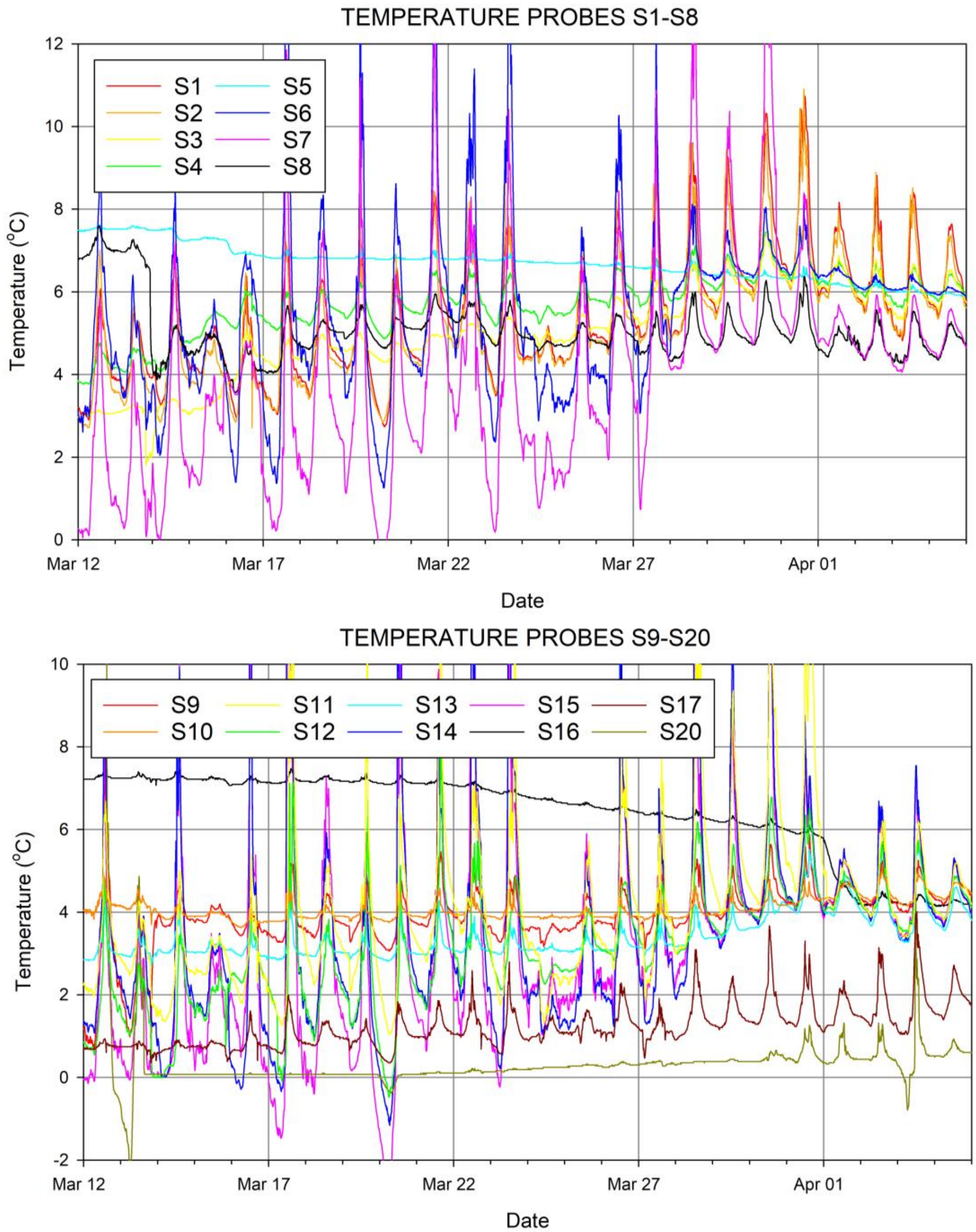

Figure 3.17 Spring temperature record for S1-S20

3.11. Supplement B

Detailed description of the summer/fall temperature data

The temperature variations at S1 were similar to S2 again on October 1st so it may have only been dry for a short period of time. S2 remained flowing/pooling and this was evident in the temperature trace and the field notes verify this. S3 transitioned from a diurnal signal range of 2°C (12°C to 14°C) to 0.75°C (close to 13.5°C) with no distinguishable diurnal pattern towards the end of July and beginning of August. Temperatures at S4 (13°C -14°C range) remained stable over the month and were similar to S3 but with a diurnal signal. S5 remained flowing and the temperature signal that was more variable than S4 or S3 and also warmer than those stations. S6 went dry on Aug 6. Prior to this it had a diurnal signal with a range of ~2°C and were similar to the S3 temperatures (13°C). After August 6 the signal range increased to 7°C and had a minimum temperature of 17°C. S7 had a very large diurnal signal (~10°C) and during August and September was warmer than any of the downstream probes with a minimum temperature of ~15°C, was dry August 3rd or 4th. S9 had a consistent daily signal (range 12.5°C to 18°C) until August 15th when the minimum temperature decreased and the maximum temperatures increased. This would suggest there was no more flow and indicates the initiation of pooling water. S11 and S14 went dry August 3rd-4th in a similar way to S6, as represented by the large increase in diurnal signal and daily minimum temperatures. S12 also went dry in early August but the temperature signal suggests it happened between Aug 7th to 14th with a slow change in daily temperature variations. Prior to August 7th S12 and S13 had similar minimum temperatures each day (12.5°C), post August 14th the lows followed the daily temperature trends and were about 4°C warmer. S13 had flow throughout August and a very stable temperature (range: 12°C to 14°C) and followed the temperature variations at S3 until September 15th when the temperature dropped 2°C and the diurnal signal was not as consistent any more. September 15th was likely the transition to pooling water. S15 went dry August 9-10, after this date the peaks were very high and it no longer had a similar trace to S2 and S11. S18 likely went dry between July 28th to August 1st (but it could have been as early as the 23rd). After August 1st the temperature variations were similar to those of SM Air (close to S5). Even before the stream at the probe went dry there were large diurnal changes in temperature and the minimum temperature (~9.5°C) was very close to the minimum air temperature, which might indicate there was not a lot of water in the channel. S19 continued to flow over the entire season with a ~5°C range in temperature each day and lows around 13°C. S20 had a similar trend to stations downstream, there was an increase in the diurnal temperature range from 3.5°C to 7°C on August 3rd, but the changes in the signal are too subtle to determine when exactly flow stopped. S10 went dry between September 7 and 12th, as inferred from the diverging signal from S8, which was located in a pool. S1 and S2 had a very similar daily temperature pattern and range (14 °C to 22 °C). S1 likely went dry on September 23 (inferred from the diverging temperature pattern at S1 and S2). See Supplement B for a more detailed description of the changes in stream temperature in summer and fall.

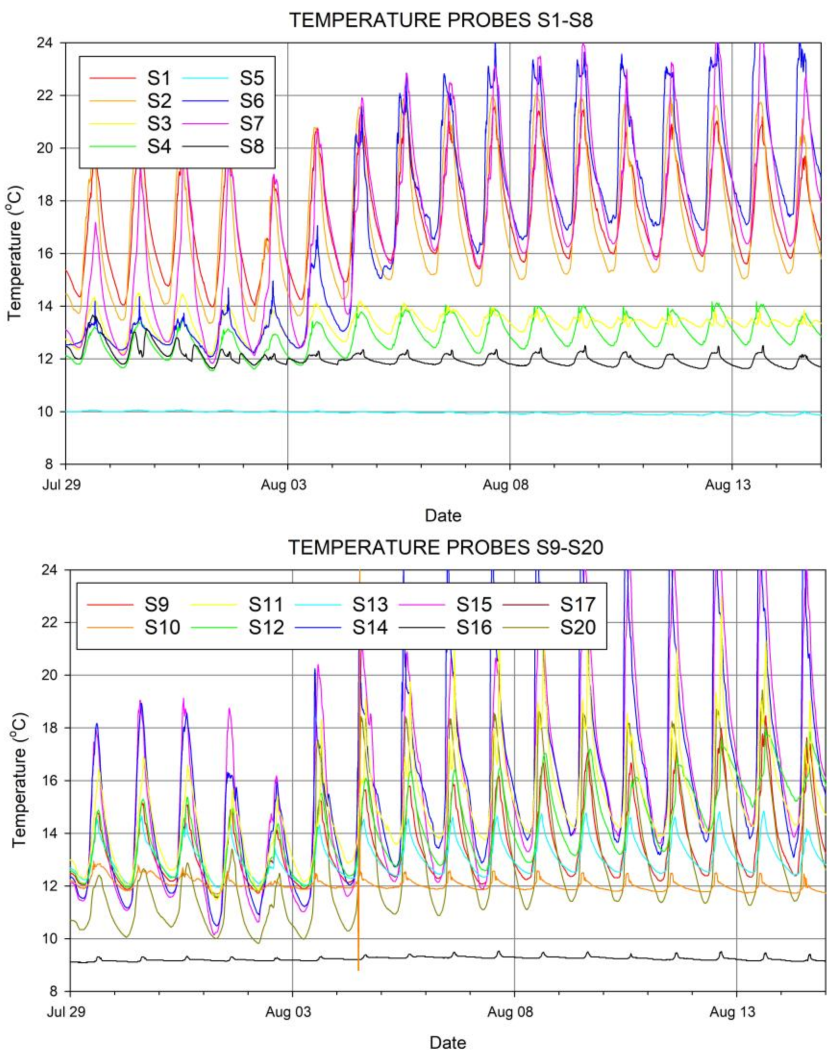

Figure 3.18 Summer temperature record for S1-S20

3.12. Supplement C

Detailed description of the temperature profile data

The LJ01 temperature profile was installed next to the piezometer (surface, 0.15 m, 0.30 m, and 0.45 m). The temperature at the LJ01 profile is at 0°C at the beginning of March. On March 3rd the 0.45 m and 0.30 m temperature started to rise. This was followed by a rise at the 0.15m sensor on March 7 and finally the surface temperature increases on March 10th. On March 10th the surface temperature probe had the largest diurnal signal and this continued throughout the month, with the amplitude decreasing with depth. All sensors followed the air temperature trends and the peaks were later deeper below the surface, indicating that was likely a losing section. There was a slow migration of the 0.15 m probe peak to the same time and value as the surface peak: this pattern is seen to a lesser extent in the 0.30 m and 0.45 m probes. The April trends continued into May. The 0.15 m and surface probes had almost the same diurnal temperature magnitude and timing. During large rain events, the 0.45 m probes was warmer than the upper probes, as would be expected. In early June there was a divergence in the pattern of the surface probe and the 0.15 m probe. The 0.15 m probe started having smaller diurnal variations and the timing of the diurnal temperature peak was later. This pattern only got stronger as the month went on. As in previous months, the diurnal temperature range was largest at the surface probe and decreased with depth.

The temperature variations seen in late June continued through July. The amplitude of the diurnal variations decreased during the month at 0.45 m, 0.30 m and 0.15 m. By the end of August, the probes at 0.30 m and 0.45 m had almost the same amplitude of diurnal temperature variations (0.3°C to 0.6°C) and a 3-hour offset in timing. The surface probe minimum temperatures were the same as air temperature and maximum temperatures only 5°C less than air temperature, suggesting the site was dry. In early September there was a reversal of the temperature gradient: the 0.45 m probe was the warmest of the three buried probes (Sept 1 -5, Sept16-22, Sept 28 onwards). The surface temperature probe continued to have the highest daily maximum temperature, until the end of October when the deepest probe was the warmest temperature.

The piezometer installed upstream of LJ22 had a temperature probe at 0.01 m and 0.38 m below the surface. The temperature time series at the upper mid piezometer start in

December 2010. From December to April there was a continual drop in temperature at 0.38 m and it was always warmer than the 0.01 m probe. The 0.01 m probe had a consistent diurnal pattern from March 3rd on. On April 1st the 0.38 m deep probe started to have a small diurnal variation (0.2°C). On April 10th the temperature at both probes was very similar each day (~2.7°C) but the maximum and minimum temperatures were higher/lower in the 0.01 m probe while the mean temperature was higher in the 0.38 m probe. The timing of the peaks was offset: the temperature at the 0.01 m probe peaked approximately 5 to 6 hours after the daily air temperature peak and the 0.38m probe another 5 to 6 hours after that. As the final snow melted, there was a gradual change (April 25 to May15) and the 0.38 m probe had consistently lower temperatures than the 0.01 m probe and the 0.01m probe started to have a very large diurnal fluctuation (5°C) each day. By May 20th the temperature at the 0.38 m probe was usually a degree lower than the daily minimum for the 0.01 m probe. The 0.38 m probe followed the overall temperature trends but had a daily variation of only 0.3°C.

After June 5th the daily variations in temperature at the 0.01m probe were very large (5°C or more). The 0.38 m probe started to have slightly larger daily variations after mid-June. From late May (30th) into early June the minimum temperatures at 0.01 m were very close to the maximums at 0.38 m and by June 14th they were often lower than the 0.38 m minimum. The diurnal maximum air temperature and the temperature at the 0.01 m probe occurred within 2 hours of each other; the 0.38 m probe peaked about 5 hours later. Field notes indicate the area went dry between June 16 and June 19, the temperature probe data indicates the area went dry on June 17th. The late June trends continued through July. There was a very small daily variation in temperature at 0.38m and a large temperature variation in the 0.01 m probe (close to air temperature).

In August the temperature at the 0.38 m probe continued to increase, rising 2.5°C over the month. In August the 0.01 m probe had slightly lower maximum temperatures than air temperature and slightly higher minimums, indicating that the site was likely dry. The 0.38m probe warmed a degree and a half over the month and the diurnal signal increased by about 0.5°C. There was a very cold day on September 1st, resulting in a 2^{o}C drop in the temperature at 0.38 m over two days. Otherwise the temperature variations were very similar to August. After Sept 7th there was a decrease in the diurnal temperature variation.

The lower mid temperature profile, was located just downstream of LJ22, and had 5 probes (surface, 0.03 m, 0.20 m, 0.35 m and 0.60 m) The lower mid temperature profile, located just downstream of LJ22, had 5 probes (surface, 0.03 m, 0.2 0m, 0.35 m and 0.60 m). In February the lower probes (0.6 0m, 0.35 m and 0.20 m) were not frozen and about 2°C while the 0.03 m and streambed probes were frozen and followed the air temperature signal closely. Temperature decreased with depth for lower three probes, and all had a ~0.1 to 0.2°C daily variation. The timing of the peaks at 0.20 m and 0.35 m was very close (within an hour) while the 0.60 m probe was about 4 hours later. Early March the two upper probes remained frozen. The 0.20 m and 0.35 m probes were consistently 0.1°C warmer than the 0.60 m probe. March 10th to 16th the air temperature remained above zero and the diurnal signal of the lower three probes almost disappeared. The upper probes continued to follow air temperature. The muted signal of the lower probes continued until two very warm days, March 30th and 31st. After March 31st there was a significant difference in amplitude between air temperature and the two upper probes, this would suggest flowing water. Starting April 1st there was a transition in the daily pattern of the 0.20 m and 0.35 m probes. They started to closely follow the 0.03 m and surface probes both in range and timing. All three peaked within 15 minutes of each other and within 0.3°C. In comparison, the 0.60 m probe had a lower daily amplitude and smoother trace, daily minimums were warmer than the upper probes by about .3°C and daily maximums were lower. The maximum temperature at 0.60 m was about 7 hours after the upper probes. These trends continued through April to mid-May.

Mid May the daily trend of the 0.60 m probe was similar to the lower portion of the signal of the upper probes. This trend continued until May 26th then the minimum temperatures at 0.60 m were similar to the probes above but the peak was still lower by at least 1°C. The timing of the peak of the upper four probes started to spread out (1.5 hours) and the signal of the 0.20 m and 0.35 m probes became much smoother. The patterns at the end of May continued into early June and by June 10th the 0.60 m daily peaks were less than the daily minimums of the probes above. June 23 the 0.60 m probe diurnal signal almost disappears (range of 0.2°C) but maintained a similar overall trend as the upper probes. The upper probes also had a significant decrease in the diurnal signal (down to 0.6°C). Early July was very similar to late June the 0.60 m probe was cooler than the upper probes and had a delayed peak with a smooth signal while the other probes all track very close together and had slightly nosier signals.

Over the course of July, the diurnal signals of the entire profile began to become more prominent, with it being most pronounced in the probes above 0.60 m. Temperatures also increased at all probes over the month, at the same rate. Aug 6th is the beginning of a pattern change. After August 6th the 0.60 m probe again had diurnal signal and varied within the maximums and minimums of the upper probes. The upper probes increased everyday – higher maximum temperatures and lower minimum temperatures. The four upper probes had very similar signals until Aug 14th then the surface probe peak increased, this was followed by another increase on the 15th and the daily maximums at 0.03 m and 0.20 m also increased; with the height of each peak decreasing with depth. On the 16th the surface, 0.03 m, 0.20 m, and 0.35 m all had a sharp decrease in the minimum. The surface was close to air temp suggesting it was dry, the 0.03 m was cooler than the 0.20 m and 0.35 m probes minimums, which fall very close together. This suggest that the 0.03 m could be dry. Starting on the 16th there was again a daily decrease in the diurnal signal of the 0.60 m probe and the mean temperature was stable at 12.2°C. August 17th none of the probes had similar signals. It is assumed the surface probe and 0.03 m were dry. The August 18th visit noted the ground was damp but no water present on surface and surface probe was dry. The 0.03 m probe had a range of ~6°C and peaked at a similar time to the surface probe (~14:00). The 0.20 m probe had a range of 1.2°C and peaked at ~20:00, the 0.35 m probe had a range of 0.3°C and peaked at ~8:30 the following day and the 0.60 m probe had a range of 0.2°C and peaked before the 0.35 m probe (~7:00 the following day). The pattern for the surface and 0.03 m probes continued to follow the air temperature patterns to the end of the month. The 0.20 m probe maintained a similar daily range and peak timing but continually increased in temperature (12.4°C to 14°C). The 0.35 m probe had almost no diurnal signal by Aug 22 and slowly increased in temperature (12.4°C to 13.5°C). The 0.60 m probe also had a decrease in the amplitude of the diurnal signal and had a mean of 12.4°C. (water level suggests it might be dry at 0.60 m Aug 17th visit).

In September the temperature at 0.60 m remained between 12°C and 12.5°C with no diurnal fluctuations, it followed the larger air temperature trends. The 0.35 m probe varied between 12°C and 13.5°C over the first half of the month and was always warmer than the 0.60 m probe. The 0.20 m probe appeared to follow the air temperature more closely and for the first five days of the month was cooler than the 0.35 m probe and other than the daily peak cooler than the 0.60 m probe as the air temperature warmed so did the probe temperature. By Sept 9th the 0.20 m probe was always warmer than the 0.60 m and 0.35 m probes. The daily variations were about 1°C and peaks happened near 20:00. The 0.03 m and surface probes appeared dry and had large diurnal signals. In the second half of September the 0.60 m probe decreased in temperature to 11.6°C and followed the larger trends in temperature. The 0.35 m probe also decreased but had larger rises and falls and followed air temperature more closely. Often the temperature at 0.35 m was cooler than at 0.60 m after Sept 17th, this was also true for the 0.20 m probe. As before, the diurnal signal was more pronounced at 0.20m than at 0.35 m. The 0.03 m and surface probes continued to appear dry (notes state dry Sept 10th).

Oct 6th to the end of recorded data (Dec 12 2011) the 0.60 m probe slowly decreased in temperature, to 5°C – followed the air temperature tends but lagged about three days. The 0.35 m probe slowly declined to 3.7°C and followed air temperature trends but lagged three days. The 0.20 m probe maintained a diurnal signal and followed the larger trends lagging air temperature by a day. The signal was not as smooth as the lower probes. It consistently had temperatures lower the 0.35 m probe but higher than the 0.03 m probe. On December 14th the temperature was 2.5°C. The 0.03 m and surface probes followed temperature until the snow arrived and then remained mostly above zero but still followed the air temperature trends. There was one exceptional date in this late fall early winter period, Nov 21 to Nov 26. There were two large snow events on Nov 12th and Nov 18th (25cm total), on Nov 22 temperatures went from -4°C to 6°C and the snow pack all but disappeared. This rapid melting caused spikes in temperature and a rewetting of the whole profile for a few days. The 0.60 m probe was at 5.8°C before the event and maintained a stable temperature through the warm event but as the air temperature dropped (~15:00) so did the 0.60 m temperature, down to 4.8°C at 18:00. The 0.35 m probe had a very different pattern. At 12:00 near the peak in air temperature the 0.35 m probe started to warm from 4.4°C to 5.1°C at 14:40, then there was a sudden drop to 2.9°C by 15:00, followed by a warming to 4.8°C at 18:00. The 0.20 m probe had no warming spike just a drop of 0.5°C between 14:40 and 15:00 then increased to 4.8°C at 18:00. The 0.03 m probe had no drop-in temperature and increased in temperature from 14:50 to 18:00 (1.6°C to 4.8°C). The surface probe was similar to the 0.03 m probe and was warm from 15:00 to 18:00, temperature increased from 0.9°C to 4.8°C. All probes remained at 4.8°C for an hour then the upper 4 probes continued to rise another

0.5°C by 20:30. A much slower rise occurred at 0.60m, starting at 20:30 can continued until 1 a.m. the following day. After 1 a.m. Nov 24 temperature started to again increase with depth. The 0.60 m probe continued to warm to 5.9°C by the 27th. The 0.35 m probe cooled 0.5°C late on the 24th and then warmed again on the 25th and stabilized close to 5°C. The 0.20 m probe cooled from noon on Nov 24th to noon on Nov 26th (5.2°C to 3.2°C). The 0.03 m probe appeared to dry by 13:00 on Nov 24th and then followed air temperature trends. The surface water appeared to have dried by 10am on the 24th and the probe closely followed air temperature again.

LJ30GW was consistently ~2.5°C from February to April 30th with no diurnal variations. The LJ01GW was constant at 0.5°C to March 4th then began increasing with a diurnal pattern. LJ01GW temperature peaked each day just after midnight, by mid-April diurnal fluctuations increased to 1°C. March 4th to March 15th the LJ01GW temperature was always less than the stream temperature, then transitioned to the mean of the stream temperature by early April. The stream temperature at LJ01 started a rapid increase from 0°C March 1st to 5°C March 11th, the diurnal fluctuation was about 2°C, the fluctuations became larger as the air temperature increased. After mid-March the daily lows in the stream were lower than the LJ01GW temperature and highs were higher. LJ25 only recorded stream temperature and this started to increase earlier than LJ01 (March 1) but did not have a significant diurnal fluctuation until March 16th. The stream temperature at LJ25 was lower than LJ30 until March 16th when the daily peaks were just higher than the temperature at LJ30, at the end of April LJ25 it had consistently higher stream temperatures than LJ30 and the daily temperature varied by approximately 2.5°C. LJ30 stream temperatures increased slowly (1°C to 2°C) from February to April 10th with almost no variation in temperature over the day. Until March 5th the stream temperature was highest at LJ30, after March 5th the temperatures were warmer at LJ01; LJ25 did not become warmer than LJ30 until March 29th (there are a few daily peaks in mid to late March that were warmer).

A diurnal temperature pattern was not prevalent until early April and was as large as 2°C by the end of April. The stream temperature at LJ30 remained cooler than the LJ30GW temperatures until April 15th, after April 15th only the daily maximum temperatures that were higher. Daily mean temperature at LJ30 and LJ30GW were about the same by April 23. Towards the end of April when all stations have large diurnal fluctuations, the daily peaks at LJ30 happen within an hour of LJ25, and about 3 hours of LJ01. The groundwater at LJ01 peaks 4 hours after the LJ01 stream temperature peaks.

In May the stream temperatures increased with distance downstream. The diurnal variations continued to get larger as air temperature increased. Stream temperature increased over the month with slight decreases during large rain events. LJ01GW continues to have a diurnal variation and equal the mean stream temperature, apart from large rain events when it had a higher temperature than the stream. The LJ30GW started to increase May 6th much like the lower station it followed the mean stream temperature apart from rain events when it was warmer. There was almost no diurnal trace in the groundwater temperatures at LJ30GW.

June temperature continued to rise and had similar trends to May. All locations had a diurnal variation except LJ30GW. LJ30GW follows the large stream temperature trends but not the daily trends. The LJ30GW continued to warm with no correlation to stream temperature. At LJ01 the LJ01GW lagged the stream temperature by about 11 hours. The daily peaks at the three stations were closer together, LJ30 and LJ25 happen within 30mins of each other while LJ01 is within an hour most days.

In July the groundwater at LJ30 continued to follow the larger trends and was slowly warming. The stream temperature at all three stations had a diurnal pattern and depended on rain and air temperature. LJ25 peaked at about 0.5°C less than the lower station and minimum temperatures were similar to the upper station. LJ01GW had a smaller diurnal signal (~1.5°C total change) and was slightly higher than the mean of the LJ01 stream. The large drops in temperature that occurred in the stream do not appear at LJ01GW.

August LJ30GW continued to increase and was close to 12°C. LJ30 temperature had a diurnal fluctuation between 11°C and 13.5°C. LJ25 had the largest daily temperature fluxes with temperatures as high as 30°C and as low as 9°C, ~16°C range each day. LJ01 varies from 14°C to 22°C, each day varying about 5°C. LJ01GW varies 2°C each day and falls between 16.5°C and 19.5°C. LJ30 only had record in early and late September. By late September the groundwater temperature was on a downward trend, there was still a diurnal pattern in the stream temperature. LJ01GW had a significantly reduced signal in early September and by September 16th almost no signal at all. LJ01 stream temperature continued to have a consistent diurnal pattern. LJ25 had the largest diurnal changes each day. All three stream stations had a very large drop in temperature during the rain event Sept 26th. After this rain event all temperatures were on a continual downward trend. LJ30GW was warmer than the maximum stream temperatures October 4th. LJ01GW was close to max daily air temperatures for most of October and by the end of October it was consistently higher than the max daily temperature and a similar temperature to LJ30GW. Although the LJ01 temperature trends followed the larger stream temperature patterns unlike LJ30 which was on a slow decrease.

Chapter 4. **Conclusions and recommendations**

This thesis focused on the streamflow characteristics of intermittent streams in the Okanagan. Intermittent streams are common throughout Canada (Buttle et al., 2012) and the rest of the world; intermittent streams constitute more than half of the global river network (Datry et al., 2014; Larned et al., 2010). However, hydrologic research has been biased to perennial streams. There has, so far, been limited research on the flow regimes of intermittent streams. Analyses of streamflow data from ten streams in the Okanagan (five perennial streams, three almost intermittent streams and two streams that went dry during some years) suggest that there are subtle differences in the flow regimes of intermittent and perennial streams beyond intermittent streams having periods of zero flow. Spring freshet tended to be earlier for the intermittent streams. Fall peak flows tended to be higher in years with large fall rain events for the intermittent streams. This led to more variability in the difference between late summer and fall streamflows for the intermittent streams. Intermittent streams also had the steepest flow duration curves and lowest master recession constants. Median spring and summer flows were not very different for intermittent streams and perennial streams. These results suggest that storage in catchments with intermittent streams is smaller than for perennial streams and that storage is depleted faster and/or there is less water available. The subtle differences also hint at shallower flow pathways. However, due to the small data set and the lack of data for truly intermittent streams, no strong conclusions about the differences in flow regimes between intermittent and perennial streams could be drawn. Currently, no intermittent streams are monitored year round in the Okanagan.

The detailed study in Long Joe creek suggested that groundwater inputs were very important for flow during the summer dry period and that the higher precipitation and snow accumulation, later melt, and lower evapotranspiration above the fog line (i.e. in the area that was in the fog in winter) caused the soil and shallow groundwater storage in the upper part of the catchment to be filled, so that it is a gaining stream for most of the spring, while the storage isn't filled in the lower part of the catchment below the fog line (i.e. not in the fog) and this is a losing stream section for most of the year, except during the very wettest part of the year. Some of the areas in Long Joe that had flow during the winter kept flowing during late summer, but others dried out. This result from

Long Joe suggests that the (shallow) groundwater storage in catchments with intermittent streams may be fairly limited and is in line with the limited shallow groundwater storage inferred from the analyses of the flow regimes for the gauging stations in the Okanagan.

Determining the groundwater inputs and losses from an intermittent catchment is difficult. The streamflow losses in the lower part of the catchment are likely due to both flow through the streambed and permanent losses to the groundwater. Reaches with coarse bed material (D84 and D50) all had periods of zero flow. The areas of zero flow expanded both from the bottom up (due to the high losses to groundwater and ET) and from the top down (due to the limited drainage area and declining groundwater levels in summer) but the dry areas never coalesced completely.

Streamflow was maintained in different parts of the catchment, signifying that there were (deeper) groundwater contributions to the stream from higher in the basin. This could be either due to topographically driven groundwater flow (also from outside the catchment (Welch et al., 2012)) or due to flow through fractures or other preferential flow pathways. However, often the flowing areas were very short and had areas of no flow upstream and downstream, suggesting that the groundwater inputs were small and quickly lost to the streambed. This high variability in the occurrence of flow suggests that the analysis of gauging station data has to be done very carefully, as the occurrence of flow and the duration of no flow at the gauging station (as well as the water yield) are unlikely to be representative for the catchment average. In order to understand the streamflow regime of intermittent streams the stream should be gauged at least at three different locations and/or the gauging station data should be complimented by at least several surveys on the occurrence and amount of flow along the stream to determine the representativeness of the gauging site.

The gauges in the Okanagan that were analysed to determine the differences in the streamflow characteristics for the intermittent and perennial streams were located at elevations that were similar to LJ07 to LJ22 (except P5 which was similar to LJ29), and therefore likely measured less flow than if they had been placed further upstream. Had Long Joe creek only been gauged at LJ25, it would have had permeant streamflow, much like P5 (which is known to be intermittent downstream of the gauging site).

The difference between early spring and late summer flows at LJ01 (no data available for LJ25 and LJ30 in March 2011) was very small. This is similar to the small difference between early spring and late summer flow at I1, I2 and A1, indicating that similar spring and summer flows might be a regime characteristic that is common to all intermittent streams in the Okanagan.

Slope perhaps is less of a predictor than Chapter 2 suggested as Long Joe creek is steeper than any of the basins studied in Chapter 2. If only the upper basin slope (above LJ25: 10.7º) is considered, the slope of Long Joe Creek is similar to the almost intermittent and P5 streams studied in Chapter 2.

Stream temperature measurements were useful to determine the source of water and the initiation of streamflow in the spring and after rainfall in fall. They were not as useful to determine the timing of the onset of the period without flow because the transition was more gradual. There were also issues with probe malfunctioning, probes being lost, buried, sitting in a pool of water or no longer in the lowest point of streambed, which reduced the data available and data quality. It is highly recommended that when temperature or EC sensors are placed on the streambed to determine the occurrence of flow, the sensors are checked regularly. Otherwise, at least for part of the sensors, the interpretation of the data with respect to the occurrence of water in the channel will be off.

This study highlights both the lack of data for intermittent streams and the difficulty in gauging intermittent streams. I recommend an expansion of the hydrometric network in the Okanagan to include streams that are intermittent. Furthermore, there needs to be a standard protocol for reporting zero flows for the national data sets (Peters et al., 2012) to be able to analyse intermittent streamflow characteristics in different regions of Canada. In the Okanagan specifically, the significance of the fog line on the spatial distribution of streamflow should be studied further to aid in the selection of streamflow station locations and to better understand the water balance of the smaller catchments in the Okanagan. Further work should also try to disentangle losses due to flow through the streambed from losses to the deeper aquifer.

4.1. References

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