Effects of climate variability on hydrological processes in a Canadian Rockies headwaters catchment

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ABSTRACT

The variability of climate in mountain headwaters has an important impact on downstream water users and extreme events such as drought and flooding. Climate strongly influences hydrological processes in mountain basins, with subsequent control on streamflow generation. By studying the variability of climate and its impact on the variability of hydrological processes, the changing interaction between climate and hydrology over time can be better understood. A meteorological and streamflow data set, from 1962 to 2013, from Marmot Creek in the Kananaskis region of the Canadian Rocky Mountains was used to calculate the key hydrological processes over a period of climate variability and change. Through the use of hydrological process modelling with the Cold Regions Hydrological Modelling platform, changes to water balance components such as precipitation, evaporation, sublimation, storage, runoff, and the processes behind these changes were investigated. Observed variability among simulated hydrological processes over time was documented. Trends in hydrological processes, climate and streamflow events and their relationships to teleconnections such as the El Niño Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO) not only documented and diagnosed historical change, but provided an indication of potential future changes to these processes and the way they will interact with a future climate. Results showed an increase trend in basin temperature and precipitation, resulting in a deeper snow pack and higher runoff volumes at higher elevations. Despite these increases, streamflow generation from the basin was unaffected due to compensation by the water budget components. Increased precipitation inputs, especially at higher elevations, were counterbalanced by losses to sublimation and evapotranspiration, especially at lower elevations, resulting in a highly resilient and consistent streamflow regime throughout the study period. Correlation analysis between the PDO and ENSO and modelled annual hydrometeorological variables and water balance component values on both hydrological year (Oct-Sep) and winter (Oct-Mar) periods revealed limited significant correlations. The Mann-Whitney U (WU) test on PDO regime shift years revealed a significant correlation with evapotranspiration in the Forest Clearing Blocks. However, the PDO and ENSO showed no consistent basin-wide impact on any hydrological process or on basin scale runoff. This research increases the knowledge of mountain basin streamflow generation's resiliency to climate change.

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This project has been fifty years in the making and would not have been possible without the blood, sweat and tears of many individuals trudging up and down the mountain collecting samples in Marmot Creek. I give them my whole hearted thanks.

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This thesis is dedicated to all of the men and women who have contributed to its fifty four year data set collected in Marmot Creek.

“I got assigned one guy who couldn’t run the snow tube- he was to stupid for that, so I had him writing down the data. We would go along and I’d call out the data, the depth and density of the snow. I stopped one place to ask him, ‘What was the last reading I gave you?’

‘I don’t know.’

‘What do you mean you don’t know? Look in your note book.’

‘Well, I didn’t write it down.’

‘Why didn’t you write it down?’

‘Well, my pencil broke.’

‘When did your pencil break?’

‘Oh, probably a couple of hours ago.’

I was so mad! I had to go back and redo the whole thing.”

-Denny Fisera

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LIST OF ABREVIATIONS

CRHM Cold Regions Hydrological Model

ENSO El Niño Southern Oscillation

GLS Generalized Least Squares

HRU Hydrological Response Unit

MCRB Marmot Creek Research Basin

MK Mann Kendall

MB Modelled Bias

NSE Nash-Sutcliffe Efficiency

PDO Pacific Decadal Oscillation

NRMSD Normalized Root Mean Square Difference

RMSD Root Mean Square Difference

SWE Snow Water Equivalent

WU Mann-Whitney U

CHAPTER 1

INTRODUCTION

This research examines the impacts of climate variability on hydrological processes in a Canadian Rocky Mountain headwater basin. Changing climate components such as rising air temperature and increased precipitation have a direct impact on the hydrological cycle in snow-dominated basins (Stewart et al., 2004; Harder et al., 2015). Changes to flow regimes are a major concern in areas of the world that are dependent on snowmelt driven natural flow such as in western Canada (Nazemi et al., 2013). Water security is a major concern moving forward into the 21st century (Wheater and Gober, 2013). The climate of western North America exhibits trends towards warmer temperatures (DeBeer et al., 2016) and recurring temperature and precipitation variations with studies showing an increase in severity and frequency of variability over the past half century (Bonfils et al. 2008a, Brown 2009). The PDO and ENSO have been linked to variability in temperature and precipitation as well as a major cause of extreme events such as drought and floods in western Canada (Shabbar and Bonsal, 2003; Shabbar et al., 2011). Hydrological resiliency, defined as retaining a similar form or function under environmental and human strain (Creed et al., 2011), of mountain basin streamflow is of great interest given recent evidence that some headwater basins in western Canada show resiliency whilst others do not (Harder et al., 2015; Dumanski et al., 2015). Determining the effect of PDO and ENSO related events on hydrological processes in the mountain headwaters of western Canada may offer insight into the observed hydrological resiliency of some headwaters (Harder et al., 2015; Whitfield and Pomeroy, 2016). The increase in climatic variability has implications for changes in mountain basin hydrology (Rasouli et al., 2014). Extensive research in the Rocky Mountain headwaters has focused on process representation using contemporary period data (MacDonald 2010; Fang et al., 2013; Hood, 2015), as well as, describing observed changes over time (Harder et al., 2015). There has been little emphasis on simulating small-scale hydrological process change due to climatic variability using reconstructed historical observation data. Understanding and quantifying the changes that have occurred in headwater catchments is required for accurate water management and improved water security.

1.1 Objectives

The goal of this study was to diagnose changes in the climate using high quality historical (1969 to 1987) and current (2005 to 2013) observational data, and the subsequent changes in hydrological processes over a mountain headwaters drainage basin. The research objectives were:

i. Determine the components of the water balance of a mountain headwaters basin using a hydrological model.
ii. Quantify the changes in hydrological processes in relation to variability in the basin's climate over a multi-decadal period.
iii. Examine the spatial variability in hydrological processes as affected by atmospheric variables such as temperature, precipitation and large scale climate indices.

To address these objectives the remainder of the thesis is divided into five chapters. Chapter 2 contains the literature review, which focuses on establishing the connection between mountain basin hydrology and climate change while exploring the use of hydrological models to quantify this connection. Chapter 3 describes the study basin, including the instrumentation and observations, corrections made to the data set, the hydrological model used for this study and the statistical analyses performed. Current and historic observational data sets were used with the Cold Regions Hydrological Model to simulate hydrological processes. The understanding of basin hydrology derived from the model served as the basis for comparing hydrological processes such as snowmelt and rainfall runoff, basin flow, intercepted and blowing snow sublimation, and evapotranspiration. Chapter 4 describes the results including streamflow analysis, significance testing for the modelled hydrological variables, quantification of the water balance components and their change over time, and the correlation of teleconnections and modelled variables. This involved examining how changes in climate may have elicited change in hydrological processes which influence streamflow. Large-scale climate indices including El Niño Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO) were also examined to help ascertain the climatic causal factors to variability in basin hydrology. Chapter 5 discusses the different significance and correlation tests that were performed and their relationships to the modelled variables while comparing the results to other studies. The goal was to determine what causes these changes in hydrological processes and the degree of influence this has on the hydrology of the basin. Chapter 6 summarizes the testing results and states the final conclusions of the study.

CHAPTER 2

LITERATURE REVIEW

2.1 Overview

Mountain hydrology is a complex system of interacting processes (Stewart et al., 2004; Mote et al., 2005; DeBeer and Pomeroy, 2009; MacDonald et al., 2010) that serves as an important source of runoff for downstream users (Vivroli et al., 2011; Nazemi, 2013). Snowmelt, in head water mountain catchments, is an important hydrological process that has shown a significant response to climate change (Adam et al., 2008; S tewart et al., 2008; Brown et al., 2009). Basin hydrology in the Canadian Rocky Mountains is well documented including rising air temperature and increased precipitation in both observational (Harder et al., 2015) and model (Pomeroy et al., 2015) studies. Teleconnections have been linked to influencing the meteorology of the Rocky Mountains and the downstream prairies (Bonsal et al., 2001; Whitfield et al. 2010). Understanding the effect of climate change on snowmelt-driven basins is paramount to the success of implementing and maintaining water security management practices in the region (Wheater and Gober, 2013). This requires quantifying the change that is occurring in these mountain basins over time and determining the link to larger climate systems to increase the understanding of the effect climate change has on snowmelt generated streamflow.

2.2 Mountain hydrology

Mountain basin hydrology receives the majority of its high elevation precipitation in the form of snow (Fang et al., 2013). Mountain basins are generally dominated by needleleaf forest cover with snowmelt being the most important annual hydrological event (Gray and Male, 1981). Needleleaf forests disrupt the timing and melt of snow by dampening turbulent energy fluxes (Harding and Pomeroy, 1996). Snow accumulation is also affected by needleleaf forests through interception. Intercepted snow is exposed to a higher rate of radiation which results in increased sublimation (Pomeroy et al., 1998) and a smaller snowpack on the ground for snowmelt (Pomeroy and Gray, 1995). Elevation is a major factor that influences temperature, phase change of precipitation and precipitation amounts in mountain basins (Storr, 1967). Snow accumulation and melt are also influenced by two major factors, slope and aspect (Pomeroy et al., 2003). Above the treeline, snow is transported by wind during which some is lost to sublimation

(MacDonald et al., 2010). In the Canadian Rockies chinook winds are responsible for above freezing air temperatures that results in mid winter melt (MacDonald et al., 2010).

2.3 Climate change

Mountain basin hydrology is susceptible to changes in climate which influences downstream water users (Harder et al., 2015; Nazemi, 2015). Understanding how the climate is changing and how this affects basin hydrology is important for future water security (Wheater and Gober, 2013). Globally, the mean surface temperature has increased approximately 0.3 to 0.6 ºC since the mid-19th century (Nichols et al., 1996). Air temperature and precipitation have seen significant increases in mountainous areas of Northwestern USA at all elevations, including an increase in rainfall as a percentage of total precipitation (Marks et al., 2010). Regionally, Bonsal and Prowse (2003) were able to identify, via observational records, temporal trends in the 0 ºC isotherm across Canada, suggesting that earlier springs were evident for the last twenty to thirty years in western Canada. Air temperature has increased in western Canada by 0.5 to 1.5 ºC from 1900 to 1998 where the greatest increase has been seen in winter daily minimum temperatures (Zhang et al., 2000). Increasing temperatures have been linked with extreme event occurrences, such as drought and floods, due to warming oceans and associated changes in weather fronts (Shabbar et al., 2010). Increasing temperatures have also been linked with long term changes in precipitation (Brown, 2000; Brown, 2009; Prowse et al., 2008; Bonsal et al., 2011). Studies have shown an increase in air temperature at the basin scale and precipitation at low elevations in Rocky Mountain headwater catchments (Valeo, 2007; Harder et al., 2015). It is important to examine these small scale basins to determine the impact this changing climate has had on hydrological processes.

Increases in precipitation match expected outcomes from Global Climate Model simulations (Akinremi and McGinn, 1998; Shook and Pomeroy, 2012). Precipitation variability and multiple day and spring precipitation events are increasing in Rocky Mountain headwater basins as well as decreases in peak seasonal snow accumulation (Harder et al., 2015). Despite these changes Harder et al. (2015) found that the volume of streamflow and timing of the peak are not changing. High elevation areas of the Rocky Mountains currently receive more than 60% of their total annual precipitation as snow (Woo et al., 2008). Precipitation phase is also closely linked to the climate through rising air temperatures. An increase in rainfall as a ratio of total precipitation could lead to further flooding in downstream regions of the prairies with an increase in summer rain on snow events resulting in faster snowpack melt rates (Pomeroy et al., 2016). Flooding in 2013 devastated large regions of southern Alberta including permanently altering the hydrology of headwater mountain basins (Fang and Pomeroy, 2016). Precipitation is also important for glacier albedo and mass balance which contributes to streamflow generation in the Rocky Mountains (Sicart et al., 2011). In runoff limited areas that are dependent on snowmelt from the Rocky Mountains, changes in precipitation amounts combined with changes in the hydrology of snowmelt dominated mountain basins would be devastating.

Increasing air temperatures have been linked with earlier snowpack melt and runoff generation which is consistent with reduced annual regular spring and early summer streamflows (Adam et al., 2009; Bonfils et al., 2008). These changes have been observed in snow depth and extent, snow cover duration, and earlier melt times (Stewart, 2008). Stewart et al. (2004) found that the shift to earlier snowmelt has been occurring since 1948 in the mountains. The effect of declining mountain snowpacks on runoff generation is an important concern for accurate water management and water security (Mote et al., 2005; Bales et al., 2006; Stewart, 2009). Many authors suggest that the cause or causes of these changes needs to be identified for the security of future water resources (Barnett, 2008; Hamlet, 2008; Nazemi et al., 2013; Wheater and Gober, 2013). Alberta rivers flowing through the prairies to the Hudson Bay have been declining over the past century (Rood et al., 2004). Changes to streamflow regime will require adaptation to the current water resource system in southern Alberta (Nazemi et al., 2013). As the primary supply of water for prairie rivers is snowmelt runoff, the Canadian Rockies snowpack is increasingly important to regions that rely upon imported water. Reliance on imported water is especially important in a region that is prone to and has shown an increase in frequency and severity of major climatic events such as drought and flood (Pomeroy et al., 2015). As these climatic events increase and grow more severe, the conditions change and expected snowmelt generated streamflow can be altered adding stress to an area that has a growing population and economy (Martz et al., 2007).

Rasouli et al. (2015) used the CRHM to predict the impact a warming climate would have on mountain basin hydrology in the US Pacific Northwest, finding that with a 5 ºC increase in air temperature and an increase in precipitation the peak seasonal snow accumulation decreased by 84% to 90%, rainfall as a fraction of total precipitation increased from 30% to 78% and snowmelt decreased by 51% to 79%. These changes also had an effect on blowing snow transport and sublimation losses from intercepted snow which decreased by a large amount (Rasouli et al., 2015). Even at just a 1 ºC increase in air temperature and a precipitation increase of less than 20% peak snow water equivalent was still decreasing significantly (Rasouli et al., 2015). These changes in snowmelt processes could increase the severity of expected changes in streamflow. Stewart et al. (2004) projected streamflow timing changes using a regression between the annual observed streamflow-timing responses of each river, local temperature and precipitation. They found that changes in streamflow, between 1995 and 2099, match the trend existing from the last fifty years, which shows an even earlier shift in snowmelt streamflow generation than has been recorded with the Rocky Mountains among those areas most affected (Stewart et al., 2004). Given the observed warming trends in the Rocky Mountains (Whitfield, 2010; Harder et al., 2015) these model prediction scenarios have drastic implications for western Canada and stress the importance of being able to quantify the change in hydrological processes to better understand how warming is effecting the current hydrology of mountain basins.

2.4 Teleconnections

The Pacific Decadal Oscillation (PDO) and El Niño Southern Oscillation (ENSO) are patterns of climate variability that have an influence on sea surface temperatures, sea level pressure and surface winds (Mantura et al., 1997). There are a few key differences between PDO and ENSO such as PDO phases last for 20 to 30 years while ENSO events typically last for 6 to 18 months, and PDO mostly influences the North Pacific but can reach down to the tropics, while ENSO is most influential over tropical areas and can extend up to the North Pacific (Zhang et al., 1997; Mantura et al., 1997).

Cyclical teleconnections, including PDO and ENSO, have been identified as influencing climate variability, particularly during the winter months (November to March), in western Canada (Bonsal et al., 2001; Bonsal and Shabbar, 2004). A positive PDO value and negative ENSO (SOI) value indicates a decrease in precipitation and an increase in air temperature as cold arctic air is prevented from entering the interior by a low pressure trough over the Pacific Ocean causing a ridge of high pressure over Western Canada (Shabbar et al., 1997; Bonsal et al., 2001). A negative PDO and positive ENSO (SOI) indicates an increase in precipitation and decrease in air temperature as cold arctic air flows into the continent interior due to a high pressure over the

Pacific Ocean facilitating a low pressure trough over Western Canada (Bonsal et al., 2001). ENSO also influences the frequency, duration, and intensity of cold (La Niña) and warm (El Ninõ) events in Canada (Shabbar and Bonsal, 2004). Phase changes in PDO and ENSO have also been linked to rising air temperature and increased precipitation amount (Whitfield et al., 2010). Of interest in this study, there was a shift in the PDO from cool to warm in 1976 and a suspected shift back to cool in the early 2000s (Bonsal et al., 2001; Whitfield, 2010). This potential shift to a cool phase could be alarming for hydrologists who have built their modeling principles around data collected largely in a warm phase which makes any insight into a study period spanning both warm and cool phases valuable.

There are studies that have looked at observational changes in relation to PDO and ENSO but there is a need to examine the correlation between simulated hydrological processes and teleconnections. Harder et al. (2015) studied the effects of teleconnections on observed hydrometeorological variables in the Rocky Mountains finding that despite a correlation between groundwater, SWE and PDO there was no correlation between PDO and streamflow which indicates a resiliency of streamflow to teleconnections. Moore and Demuth (2001) studied the effects of climate variability on streamflow generated by Place Glacier located in the coastal mountains of British Columbia. They found that the net balance of the glacier was directly related to the PDO, which had shifted in 1976 from a cold to warm phase. Also, earlier springs linked with teleconnections have been found in major northward flowing river systems in Canada including the Mackenzie River (Prowse et al., 2010)

2.5 Hydrological Modelling

Hydrological models are used to simulate hydrological processes using energy and precipitation inputs. The complexity of hydrological models, their required inputs and parameters, and the processes simulated varies. Grayson and Blöschl (2001) define three model characteristics including the type of algorithm, input and parameters method, and the spatial representation. The type of algorithm refers to whether a model is empirical, conceptual or physics based, the input parameters are either deterministic or stochastic and, the spatial representation is either distributed or lumped (Grayson and Blöschl, 2001).

There are a number of models that are suitable for hydrological modeling in a cold regions climate. Environment Canada developed a coupled land surface and hydrological model known as the Modelisation Environmentale Communitaire (MEC) – Surface and Hydrology (MESH) (Pietroniro et al., 2007). The MESH model is expanded from the MEC which created an environment where different land surface models could coexist within a shared modelling framework to allow for the same grid, time step, interpolations and output specifications to be easily compared (Pietroniro et al., 2007). With the Grouped Response Unit (GRU) approach, the MESH can set parameters for each landscape class to facilitate the calibration of the model on the whole domain versus at the sub-basin level (Pietroniro et al., 2007). In contrast, the Hydrological Response Unit (HRU) assumes that the study area can be broken into units corresponding with variations in land cover, slope, aspect, soils and elevation (Fluegel, 1995). The key difference between GRUs and HRUs is that the HRU can transfer mass from HRU to HRU (ie: From an alpine HRU to a treeline HRU in a mountain basin), while GRU can only move mass around among land cover types in their determined grid. RAVEN is a semi-distributed HRU based hydrological model. RAVEN uses empirical relationships to simulate cold-regions processes, such as using temperature index model to calculate snowmelt (RAVEN User's and Development Manual). The temperature index model is not practical for a climate change study as it does not consider the individual energy fluxes, also a climate change study can not use calibrated melt factors due to non-stationary.

Substantial hydrological modelling has been performed in western Canada and particularly in the Rocky Mountains looking at numerous processes including canopy interception, Chinook winds, snowmelt and runoff generation (Ellis, 2010; MacDonald, 2010; DeBeer and Pomeroy 2010; Fang et al. 2013). For this study, the Cold Regions Hydrological Model (CRHM) was chosen for its physics based approach with a semi-distributed vegetation representation (Fang et al., 2013). The CRHM also has an extensive background in modelling snow hydrology in mountain basins while producing outputs that are comparable to observed measurements (Pomeroy et al., 2007; Fang et al., 2013). Comparing observed and modelled results showed that the greatest variation in potential snowmelt energy from sub-canopy shortwave irradiance, in response to a change in cloudiness and forest-cover density, will occur under south-facing forests (Ellis and Pomeroy, 2007). Snow accumulation and melt simulations were compared to observed data in both forested and clearing sites at different elevations in a

Rocky Mountain basin, the CRHM was shown to be able to exhibit the variation in snow accumulation between forest and forest clearing sites but slightly over estimated sublimation losses from the canopy (Ellis et al., 2010). Pomeroy et al. (2012) performed a forest cover disturbance simulation of a mountain basin affected by mountain pine beetle finding that in all cases the impact on basin hydrology was small with an increase in snowmelt below 10% and of streamflow less than 2%. The low percentages of disturbance can be attributed to the forest cover affected by mountain pine beetle, lodgepole pine, being found mainly at lower elevations where precipitation amounts are much lower than higher elevations (Pomeroy et al., 2012). Disturbances at higher elevations, such as forest fire, exhibited up to an 8% increase in spring and summer streamflow as a result of increased snowmelt and sublimation of winter snow, but the largest response to forest disturbance came from peak daily streamflow charges with an increase of almost 25% (Pomeroy et al., 2012).

Sublimation losses and blowing snow sublimation losses were found to be significant in the low alpine zone of a Rocky Mountains basin using the prairie blowing snow model (PBSM) and a snowpack ablation model (SNOBAL), with a snow mass loss of 20% to 32% as a percentage of cumulative snowfall (MacDonald et al., 2010). Through the use of a physically based snowmelt energy balance model (SNOBAL), DeBeer and Pomeroy (2009) were able to simulate snowmelt using point observations of SWE from a nearby meteorological station finding that pre-melt distributions of SWE and spring melt rates display a large amount of spatial variability in slope within an alpine cirque. A negative association was found between daily melt rates and SWE in the early melt period when comparing modelled results (CRHM) with measurements, with the deeper snow packs requiring more energy to initiate melt (DeBeer and Pomeroy, 2010). Precipitation-phase partitioning methods (PPMs) used for simulating cold-region hydrological processes were examined by Harder and Pomeroy (2014) and found reduced uncertainty when using physically based PPMs for snowpack prediction in mountain basins. Simulating hydrological processes over a range of temperature thresholds quantified the uncertainty, reaching 20% for the rainfall fraction, 0.4 mm/day for basin discharge and 160mm of peak snow water equivalent (Harder and Pomeroy, 2014). Uncertainties in snow modelling due to variance in mathematical representations of windflow were examined by Musselman et al. (2015) who found that the snow mass budget simulated by a snow model was highly sensitive to windflow with a difference in cumulative season sublimation ranging from 10.5% to 19% of seasonal snowfall depending on the windflow model used.

Modelling of Spanish Pyrenees mountain hydrology with the CRHM by Lopez-Moreno et al. (2013) reported that variability in snow accumulation and duration of snowpack was influenced heavily by the slope. There was a large response in sensitivity of the snowpack to warming on south facing slopes versus north facing slopes. Two snowmelt-dominated mountain basins in western China were modelled by the CRHM to simulate hydrological processes, the model performed well in reproducing measured streamflow with a r^2 of 0.83 and a Nash-Sutcliffe coefficient (NSE) of 0.76 (Zhou et al., 2013) when energy balance snowmelt and infiltration to frozen soils routines were enabled. To increase the understanding of streamflow response to changes in mountain basin hydrology for one of the largest rivers in Chilean Patagonia, the CRHM was used to simulate hydrological processes in the upper Baker River basin (Krogh et al., 2015). Observed precipitation was found to have almost no predictive power, with a NSE less than 0.3, if used to force the hydrological model, but by using atmospheric model reanalysis data the NSE increased to greater than 0.7 (Krogh et al., 2015).

2.6 Summary

Mountain basin derived streamflow is an important resource for both economical and natural systems (Shook and Pomeroy, 2012; Wheater and Gober, 2013). Hydrological processes responsible for Rocky Mountain generated streamflow are being influenced by anthropogenic induced climate change, resulting in rising air temperatures and increased precipitation amounts (Harder et al., 2015), increasing occurrences of downstream flooding (Shook, 2015; Shook and Pomeroy, 2015) and changes in flood peak (Whitfield and Pomeroy, 2016). As the temperature warms and precipitation amounts and frequency are altered, rain on snow events will become more common which accelerates early summer snowmelt rates and further alters the streamflow (Pomeroy et al., 2016). Assessing the trends behind changes in the hydrology of the Rocky Mountains is an important step in identifying the subsequent changes to basin hydrology. There has been a lot of emphasis on how hydrological processes are changing and what is causing this change, but the amount of change needs to be quantified and examined. This study will use a unique historical observational data set and process model to diagnose these changes in hydrological processes at a local scale in an important headwater catchment. The opportunity to examine these hydrological processes closely from a historical perspective is a current gap in the literature and will offer insight into how much change is occurring as well as the resiliency of mountain basin streamflow.

CHAPTER 3

METHODS

3.1 Marmot Creek Research Basin

The location of this study is the Marmot Creek Research Basin (MCRB), located on the eastern slopes of the Front Ranges of the central Canadian Rockies. The basin is situated west of Calgary in the Kananaskis Valley, adjacent to Nakiska Ski Resort (Figure 3.1). The MCRB ranges in elevation from 1590 m (above sea level) to 2800m and its total area is 9.4 km^2. The basin consists of three sub-basins; Cabin Creek (2.35 km^2), Middle Creek (2.94 km^2) and Twin Creek (2.79 km^2) and a confluence sub-basin (Fang et al., 2013). With studies founded in 1962 by a variety of federal and provincial agencies, this research basin is unique because it is one of few long term hydrology research sites in the Canadian Rocky Mountains (Rothwell et al., 2016). The amount of research that has been conducted in this basin, paired with studies conducted in similar sites around the world, make it an ideal location for examining the effects of climate change on hydrological processes in a headwaters catchment.

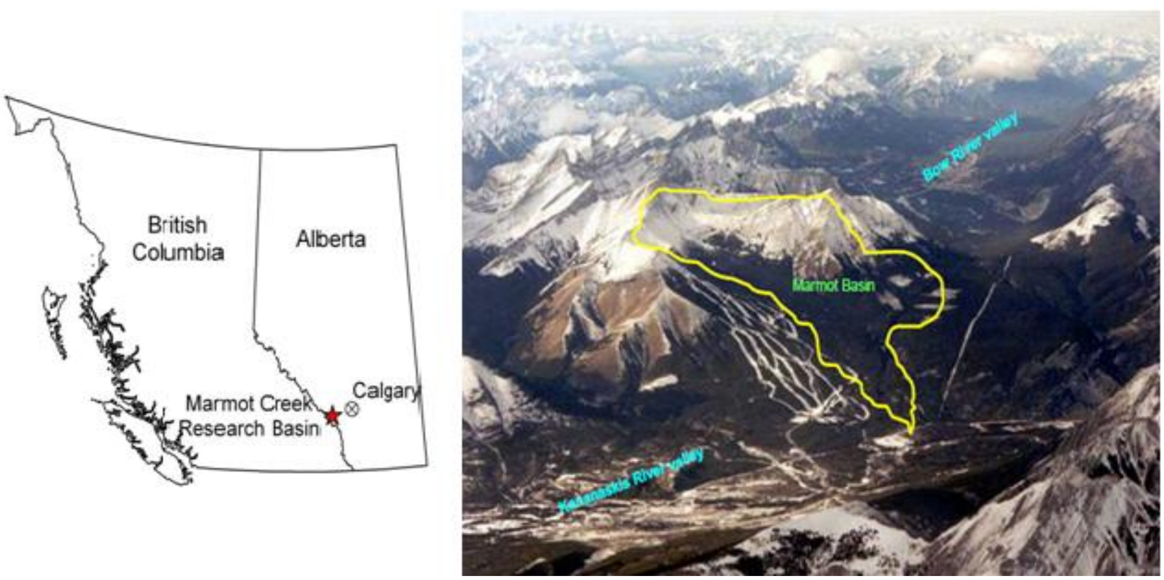

Figure 3.1: Location of Marmot Creek Research Basin (MCRB) (outlined in yellow) in the Canadian Rocky Mountains. Location in Western Canada denoted on the left, outline of basin adjacent to Nakiska Ski Hill on the right. (Figure courtesy of the Centre for Hydrology, University of Saskatchewan).

The majority of vegetation in MCRB is needleleaf forest with Engelmann spruce and subalpine fir in the higher elevations and lodgepole pine in the lower elevations (Fang et al., 2013). The vegetation thins to alpine larch and short shrubs at the treeline, with exposed rock making in the upper alpine portion of the basin (Figure 3.2). There are clear-cut honey comb patterns present in the forested areas as part of a study to determine the effects of clear cutting on water availability in the 1970s and 1980s (Golding and Swanson, 1978). The soil in the basin freezes seasonally and is coarse and permeable which allows rainfall to infiltrate rapidly (Jeffrey, 1965). MCRB's weather is influenced by continental air masses that result in long cold winters and cool wet springs. Chinook winds allow for brief periods of above freezing air temperature during winter and produce strong winds (MacDonald et al., 2010). The annual precipitation changes with elevation from 600 mm at the lowest elevation of the basin to 1100 mm at the highest elevation (Storr, 1976). Approximately 70% to 75% of the precipitation occurs as snowfall, increasing from lower to higher elevations (Fang et al., 2013).

3.2 Basin Instrumentation and Observations

Hydrometeorological instrumentation in MCRB is described in two sections, the historical sites and current sites:

The historical sites, including Confluence 5 (50.947, -115.204), Cabin 5 (50.975, -115.182), and Twin 1 (50.9569, -115.204), were established in the early 1960s by the Meteorological Service of Canada and Canadian Forestry Service, with data deemed useable for this study from October 1st 1969 to September 30th 1987. These dates were chosen because of their availability of data, particularly wind speed observations, acquired from the Canadian Forestry Service and Environment Canada. Confluence 5 is a mid-range forest clearing site with an elevation of 1753 m, Cabin 5 is a high elevation forested site with an elevation of 2170 m, and Twin 1 is an alpine site bordering the treeline with an elevation of 2286 m. The historical sites recorded hourly temperature, relative humidity, and wind speed and daily precipitation. Twin 1 recorded daily wind speed averages. Further instrumentation that supplied data for this study included the University of Calgary Biogeosciences Institute Barrier Lake Field Station (BGSI) (51.027, -115.035) located approximately 13 kilometers north-east of the MCRB at an elevation of 1391 m.

The current observation sites, including Upper Clearing (50.9565, -115.175), Vista View (50.971, -115.179), and Fisera Ridge (50.9568, -115.204), were established by the University of Saskatchewan, Centre for Hydrology, in 2004 with data useable for this study from October 1st 2005 to September 30th 2013. These dates were chosen because of the availability of a meteorological data set that was collected and maintained by the Centre for Hydrology and tested through the use of a hydrological model in the basin. Upper Clearing is a mid-range forest clearing site with an elevation of 1845 m, Vista View is a high elevation forested site with an elevation of 1956 m, Fisera Ridge is an alpine site bordering the treeline with an elevation of 2325 m. Data were also drawn from additional modern sites at the ridge line (Centennial Ridge (50.994, -115.193)) and valley bottom (Hay Meadow (50.9441, -115.138)). The current sites recorded hourly temperature, relative humidity, wind speed, precipitation, and incoming short wave solar radiation.

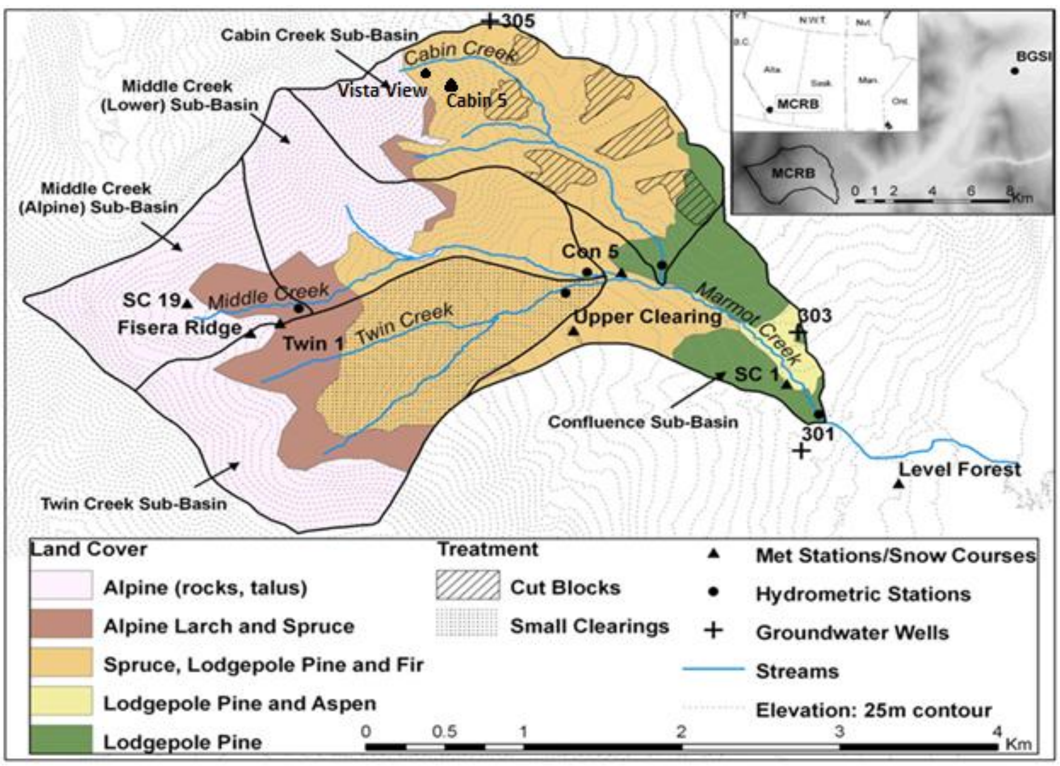

Figure 3.2: Land cover characteristics of Marmot Creek Research Basin (MCRB) (outlined in black) and the associated sub-basins. The triangles represent meteorological stations including the historic sites (Twin 1, Confluence 5 and Cabin 5) and current sites (Fisera Ridge, Upper Clearing and Vista View). (From Harder et al., 2015, reproduced with permission of John Wiley and Sons Ltd. Copyright © 2015)

The Water Survey of Canada operated a streamflow gauge at the outlet of Marmot Creek year round from 1962 to 1986 and then seasonally (May-October) until June of 2013 when flooding occurred which has since diverted creek flow from the weir and rendered it useless. There is a 17 year gap in the data between 1987 and 2005 that is addressed in section 3.3.

3.3 Data Set Correction

Observational data are actual recorded measurements where as reanalysis data is assimilated using numerical schemes for a given time period. This study is relatively unique because it used a fully reconstructed observational data set from multiple observations in a mountain basin. Reanalysis data also runs into errors with spatial and temporal interpolation with the numerical schemes. To ensure the quality of an observational data set, quality assurance is performed. The meteorological data were collected from the listed historical sites in section 3.2 which consists of hourly measurements of air temperature, relative humidity and wind speed as well as daily total precipitation and daily average wind speeds. The majority of the data were already digitized, but portions were manually entered, mainly the daily wind speed averages, from scanned hard copies from the Canadian Forestry Service of Natural Resources Canada in Edmonton, Alberta. There were a number of steps for quality assurance with the data, including removing inconsistencies, such as sudden large increases in measurements, and outliers. Table 3.1 shows the thresholds used to identify outliers and the rate of change (ROC) used to identify inconsistencies in the data which were applied to air temperature, relative humidity and wind speed for the historic period.

Data were infilled for the historic period using simple linear regression equations developed between the historic sites. The inverse weighted method was not used because it places importance on spatial distance assuming points that are close together are more alike than those farther apart. This is not ideal for a small basin where data variability in the observed meteorological conditions was primarily due to differences in elevation between stations, as opposed to spatial distance. Data were infilled using the Water Information Systems KISTERS software (WISKI) and R with the following procedure. If a gap in the data was three hours or less, then the average of the hour preceding the gap and the hour following the gap was used. If the gap was longer than three hours then a regression equation was used. A “use best option” script was written through the WISKI modelling program to select the site with the greatest coefficient of determination t (R^2) to use for infilling. For example, if Twin 1 was missing four hours of air temperature then the site with the coefficient of determination, in this case Cabin 5, was used to infill the data with the appropriate regression equation. If recorded data were not available for Cabin 5 for this date and time it would then go to option two which was Confluence 5 and down the list until it found a match. The order of selection would be historical sites ranked on greatest to lowest coefficient of determination then BGSI, and if none of these options were suitable, calculated long term averages were used.

Table 3.1: Threshold and rate of change limits for each variable applied for quality assurance. Air temperature, relative humidity, wind speed and precipitation (daily) were used for the historic period and air temperature, relative humidity, wind speed, precipitation (hourly), shortwave radiation and soil surface temperature were used for the current period.

Variable	Unit	Maximum	Minimum	ROC Limit
Air Temperature	C	40	-40	10
Precipitation	mm	30 (hourly)	0	N/A
Relative Humidity	%	100	0	30%
Wind Speed	m/s^{-1}	40	0	N/A
Shortwave Radiation	W/m^{-2}	1368	0	1368
Soil Surface Temperature	C	50	-40	10

Hourly air temperature was the most complete portion of the data set, as shown in Figure 3.3, and required very little infilling compared to other meteorological variables. When necessary, it was regressed from adjacent sites using the WISKI script previously mentioned and used a constant environmental lapse rate of 0.75 ºC per 100 meters (Fang et al., 2013). Relative humidity (RH) was converted to vapour pressure to remove any influence of temperature at the current station. It was then converted back to RH based on the appropriate temperature from its regressed station. Figure 3.4 is an example of relative humidity data that required threshold corrections and gap infilling from the Twin 1 site.

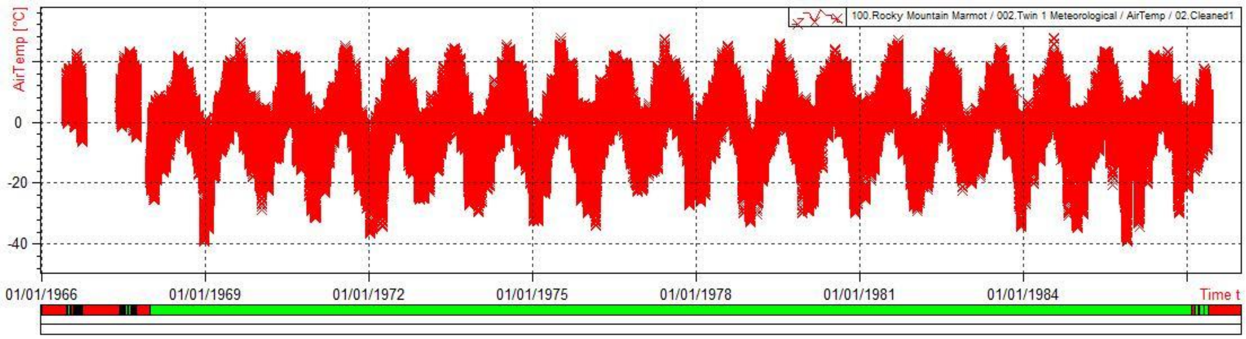

Figure 3.3: Air temperature measurements for Twin 1 over the historic time period. The x axis displays the date (1966 to 1987) and the y axis displays air temperature in Celsius (-50 to 40), red X's represent the actual measurements. The bar on the bottom displays gaps in the data, green representing available measurements, black represents individual missing values and red represents consecutive missing values.

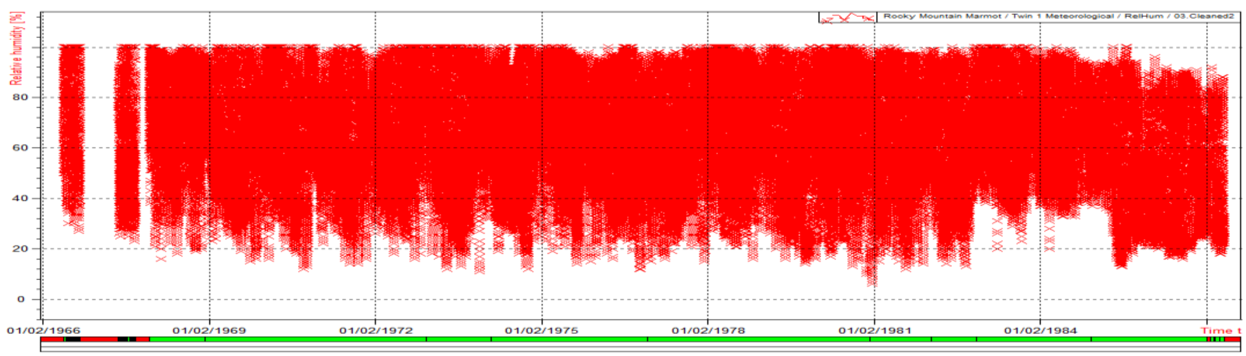

Figure 3.4: Relative humidity measurements for Twin 1 over the historic time period. The x axis displays the date (1966 to 1987) and the y axis displays relative humidity (%), red X's represent the actual measurements. The bar on the bottom displays gaps in the data, green representing available measurements, black represents individual missing values and red represents consecutive missing values.

Gaps in precipitation were filled with an equation that uses a seasonal elevation gradient to interpolate two stations at different elevations – this gradient was measured from precipitation gauges at multiple elevations in MCRB (Pomeroy et al., 2011):

$$p_B = p_A + \frac{p_A(Elev_B - Elev_A)}{1000(gradient)} \tag{3.1}$$

where A and B represent the two stations being interpolated, p is the precipitation amount in mm, Elev is the elevation of the station and the gradient varies seasonally (September to October 0.0594, November to March 0.0413, April to May 0.1002 and June to August 0.0988). Figure 3.5 is an example of daily precipitation data from Confluence 5 over the historic period that required infilling.

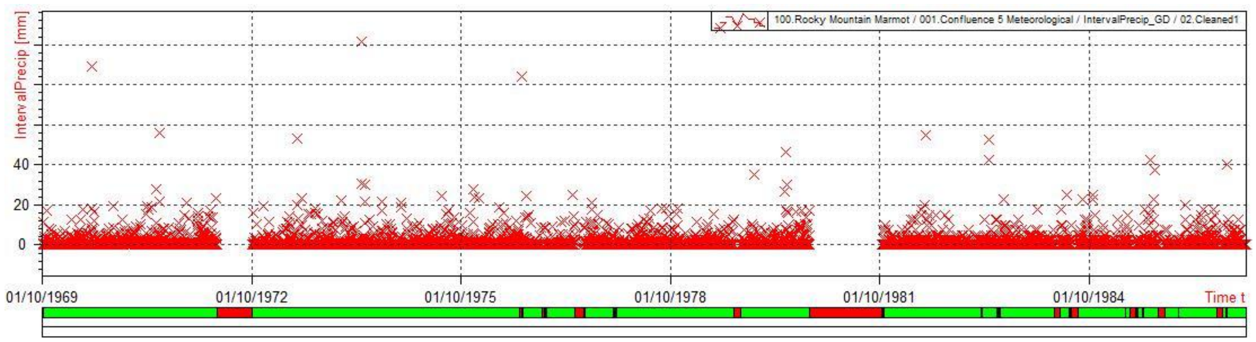

Figure 3.5: Daily total precipitation for Confluence 5 over the historic time period. The x axis displays the date (1966 to 1987) and the y axis displays daily total precipitation (mm), red X's represent the actual measurements. The bar on the bottom displays gaps in the data, green representing available measurements, black represents individual missing values and red represents consecutive missing values.

Wind speed for the historic period was the most challenging meteorological variable to infill due to sparse availability of hourly data and many years having only daily averages. Figure 3.6 shows the condition of the raw wind speed data for Twin 1 and Confluence 5 in the 1980s before quality assurance. The wind speed data were broken up into two categories, daily wind speed averages that were recorded at Twin 1 from 1969 to 1980 and hourly wind speed measurements that were recorded at all three historic sites from 1981 to 1987. Daily average wind speeds were not suitable for this study since the hydrological model requires hourly wind speeds as its blowing snow components were calibrated to observations and wind speeds made at sub-hourly intervals (Pomeroy et al., 1993).

There were multiple steps required to manipulate the daily measurements into hourly measurements for the purpose of modelling. The first was to determine a regression equation for the 1980's data that could be used in the 1970s to extrapolate the Twin 1 data to the other historic sites. This was done by averaging the hourly data to daily averages and then testing its correlation from Twin 1 to Cabin 5 and Confluence 5 and developing a regression equation. After this equation was established, it could be used to provide wind data for the other historic sites in the 1970s. The second step was to determine how much effect temporal scaling from daily to hourly wind speeds would have on the model results. Test runs were done of the various modelled components and it was deemed to have an important effect on wind sensitive blowing snow processes; blowing snow transport and sublimation. This determined that the daily wind speed was not sufficient to use for modelling and would need further correction.

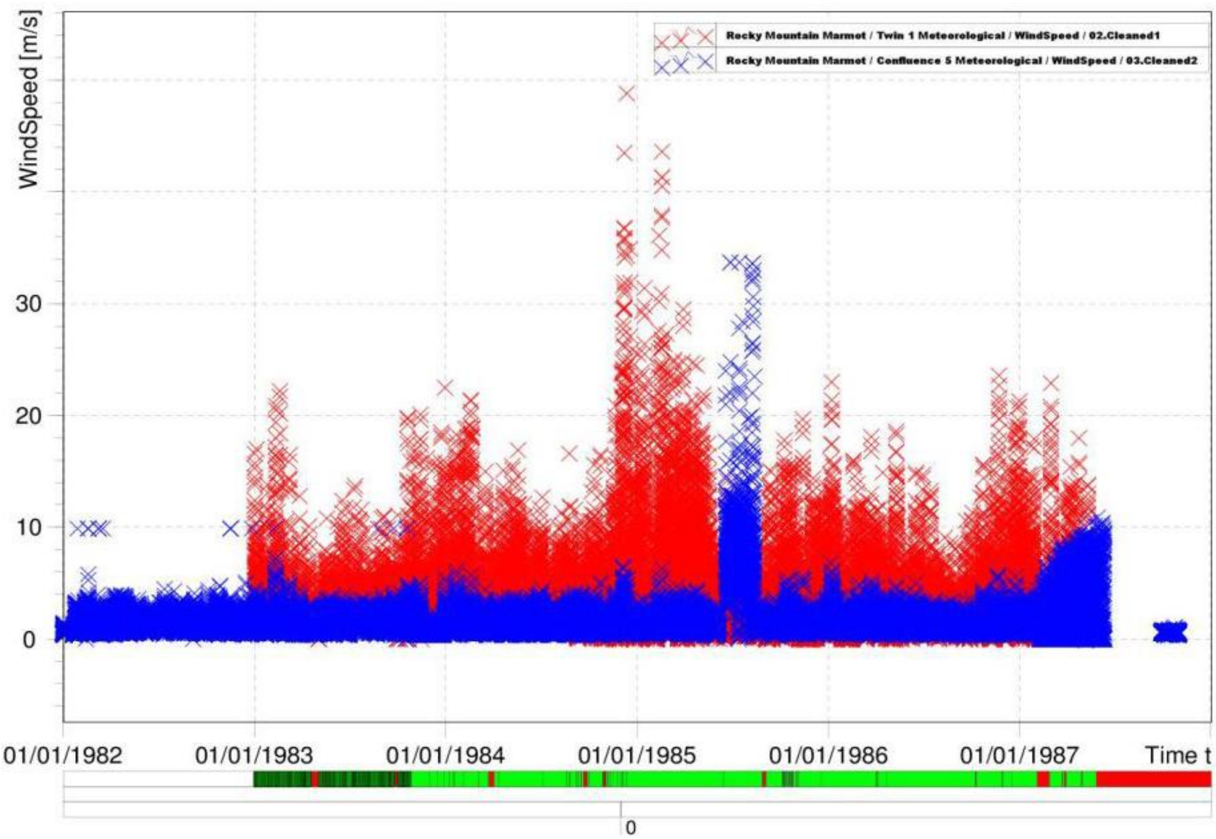

Figure 3.6: Hourly wind speed measurements (m/s) for the historic period. The x axis displays the date (1982 to 1987) and the y axis displays hourly wind speed measurements (m/s), red X's represent measurements from Twin 1 and blue X's represent measurements from Confluence 5. The bar on the bottom displays gaps in the data, green representing available measurements and red represents missing values.

Various factors were applied to the wind speed in an attempt to see if increasing the wind speed by a certain percentage would prove sufficient for correcting it to comparable hourly values. The factors showed improvement on the modeled results, however the modelled values for the historical and current period did not compare. It was determined that the poor model performance for the historical period was caused by a lack of wind speed from a ridgeline station in the historical data set to be contrasted against the wind speed from the Centennial Ridge station used in the current data set. Two factors, shown in Figure 3.7, were determined by calculating the accumulative wind speed at the Centennial ridgeline station for both historical time periods. These factors were then used to interpolate to a point to adjust the wind speeds in the 1970s and 1980s to make the data comparable to the modern period. The model was then tested with these values and produced model runs that were acceptable as they were comparable to model outputs from the current period.

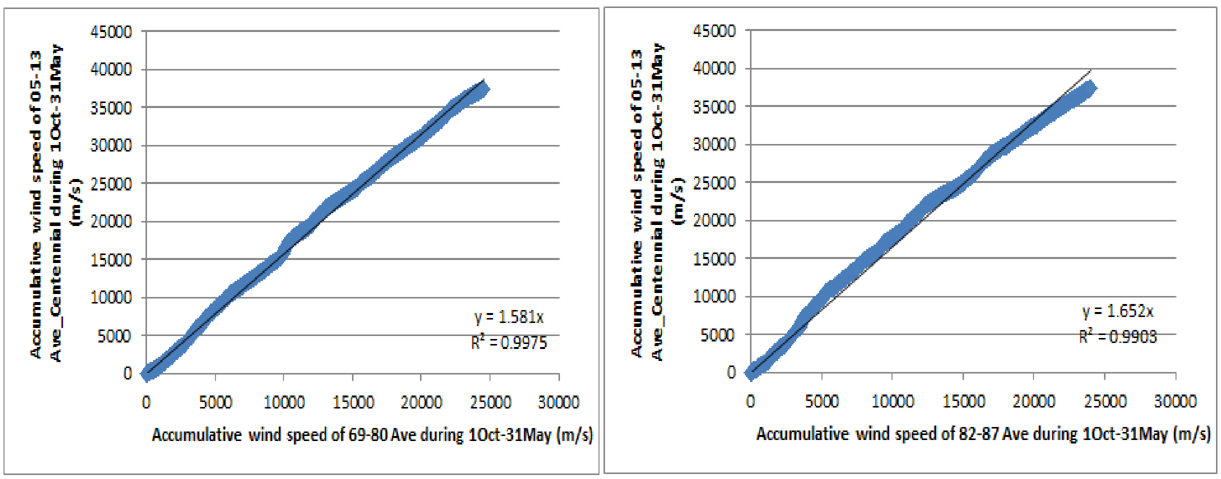

Figure 3.7: Cumulative wind speed correlation between historical data and current data for Centennial Ridge (left) and Twin 1 (right). The x axis shows the cumulative wind speed for both historic periods (1969 to 1980 and 1982 to 1987) at Twin 1. The y axis displays the accumulative wind speed for Centennial ridge from 2005-2013. The blue plots indicate the measurements and the black line is the linear regression between the two stations.

The historical data did not include incoming shortwave radiation, which is used to calculate evapotranspiration, soil thaw, snowmelt and various other hydrological processes, so the Annandale method, which is an empirical method based on the range in daily air temperature, was used to estimate solar radiation (Annandale, 2001):

$$\tau_D = k_{RS} \, (1 + 2.7 * 10^{-5} \, Alt\Delta T^{-5}) \qquad (3.2)$$

where k_{RS} is the adjustment coefficient (0.16 interior locations and 0.19 for coastal regions), Alt is the site altitude in meters and ΔT is the change in temperature. Annandale is considered to be superior to alternative methods for estimating solar radiation because the effects of altitude on transmittance are integrated into the equation and it does not require external factors to be calculated (Shook and Pomeroy, 2011). However, Anandale calculates a daily value of incoming solar radiation and many hydrological models require hourly values to simulate processes. Shook and Pomeroy (2011) developed a modified version of the Annandale method that could interpolate hourly values from daily estimates using an equation that requires simply the date and the hour:

$$R(h,d) = max\left[\sin\left(\frac{c1\pi h}{24} + c2\right)c3, 0\right] \tag{3.3}$$

where h is the hour, d is the date, c1, c2 and c3 are functions of day number. The functions of day numbers are calculated as follows:

$$c3 = 8.840749 * 10^{-5}\, d^2 - 0.030195\, d + 5.33384 \tag{3.4}$$

$$c2 = -0.963832\, c3 + 7.86346 \tag{3.5}$$

$$c1 = -0.658542\, c2 + 5.18605 \tag{3.6}$$

The data for the current period (2005 to 2013) used the thresholds and rates of change outlined in Table 3.1 for air temperature, relative humidity, wind speed, precipitation (hourly), shortwave radiation and soil surface temperature. Data were infilled for the current period using regression equations between sites, using the site with the greatest coefficient of determination. Data gaps were infilled using the same method with the historic data; if a gap in the data was three hours or less, then the average of the hour preceding the gap and the hour following the gap was used. If the gap was longer than three hours then a regression equation was used. When regressing air temperature, a constant environmental lapse rate of 0.75 ºC per 100 m was used (Fang et al., 2013) and gaps in precipitation were filled using the same method as the historic data outlined in Equation 5, a seasonal elevation gradient to interpolate two stations at different elevations (Pomeroy et al., 2011). Data in the current period were considered to be higher quality in terms of collection methods and consistency than the historic period and did not require extensive alterations for use in this study. The gap in data from 1987 to 2005 was not infilled due to a lack of available hourly data in the area during this time. Daily measurements were taken at sites such as BGSI, but nothing was recorded at the same level of detail as the Marmot Creek Research Basin and reanalysis dataset use was avoided to restrain uncertainty.

3.4 Cold Regions Hydrological Model (CRHM)

For the purpose of modeling in this research, a model was needed that had exhibited the ability to simulate complex hydrological processes in a cold regions mountain basin environment with respect to varying land cover types and elevation gradients. The Cold Regions Hydrological Model (CRHM) is an object oriented, modular modelling platform used to assemble physically based hydrological models that was developed at the Centre for Hydrology, University of Saskatchewan (Pomeroy et al., 2007). The model uses meteorological observations to simulate hydrological processes such as blowing snow, infiltration to frozen soils, evapotranspiration, sublimation, snowmelt, and interception.

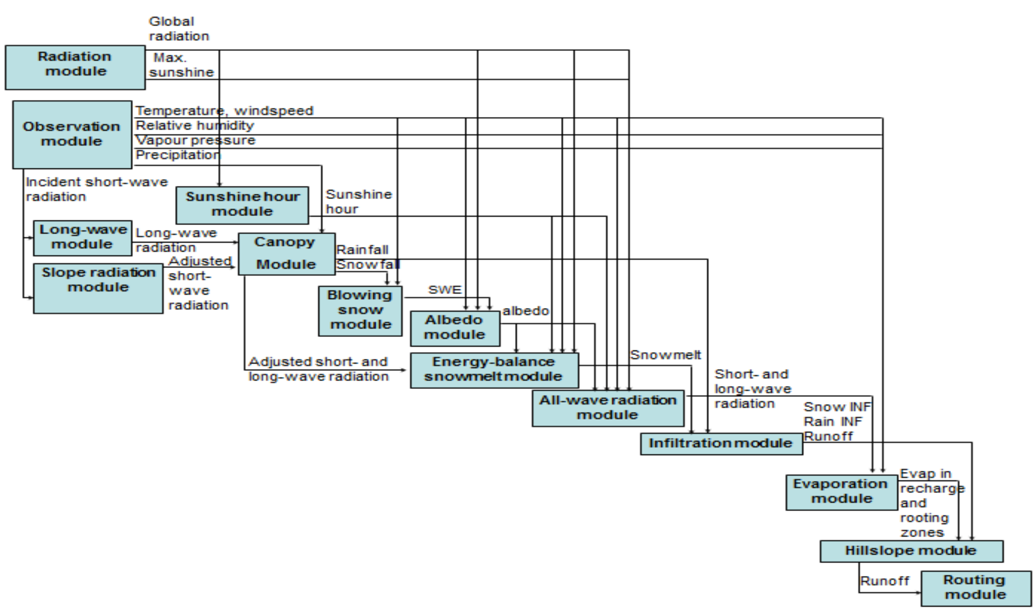

Figure 3.8: Flowchart showing the modules used to simulate hydrological processes with the Cold Regions Hydrological Model platform. Model execution is from left to right.

The CRHM allows for customization of the model through its use of parameters to represent data uncertainty or unavailability. The CRHM uses basin specific spatial configurations, spatial resolutions and physical process modules to model hydrological process within the basin. Hydrological response units (HRUs) are used to represent the physical characteristics of the landscape, through spatial units of mass and energy balance calculations, and may vary based on the level of physical complexity chosen for the simulation (Fang et al., 2013). HRUs are landscape units that are grouped based on similar soil types, vegetation and topography. For this study the HRU groups analyzed were Alpine, Treeline (transition zone between Alpine and Upper Forest), Upper Forest, Forest Clearing Blocks, Lower Forest, and Basin.

The CRHM and the associated modules were chosen for this study based on their extensive use within the MCRB and proven ability to reproduce observation quality modelled results (Pomeroy et al. 2007, Ellis et al. 2010 and Fang et al., 2013). The modules used to model hydrological processes in Marmot Creek are shown in Figure 3.8 and are listed with a brief description below:

- The observation module reads the meteorological data from an output file while adjusting temperature for environmental lapse rate, precipitation for elevation, and wind induced undercatch while supplying these inputs to other modules.
- The radiation module uses the latitude, elevation, ground slope, and azimuth to calculate the theoretical global radiation, direct and diffuse solar radiation and sunshine hours (Garnier and Ohmura, 1970). This provides radiation inputs to the sunshine hour module, the energy-budget snowmelt muddle and the net all-wave radiation module.
- The sunshine hour module uses incoming shortwave radiation and maximum sunshine hours to estimate sunshine hours which generates inputs to the energy-balance snowmelt module and the net all-wave radiation module.
- The slope radiation module uses a measurement of incoming shortwave radiation on a level surface to estimate incident shortwave to a slope.

- The longwave radiation module uses measured shortwave radiation to estimate the incoming longwave radiation (Sicart et al., 2006) which is then put into the energy balance snowmelt module.
- The albedo module was used to estimate the snow albedo for the winter and into the melt period (Verseghy, 1991). It is also used to indicate the beginning of the melt period for the energy balance snowmelt module.
- The canopy module was used to calculate shortwave and longwave sub-canopy radiation and estimate the snowfall and rainfall that is intercepted by the forest canopy (Ellis et al., 2010). There are multiple options involved with this module including open environment, small forest clearing environment, and forest environment.
- The blowing snow module was used during the winter period to simulate the inter-HRU wind redistribution of snow transport and sublimation loss (Pomeroy and Li, 2000).
- The energy-balance snowmelt module is a version of the SNOBAL model that was developed to simulate the mass and energy balance of mountain snowpacks (Marks et al., 1998). This module calculates the energy balance of radiation, sensible heat, latent heat, ground heat, advection from rainfall, and the change in internal energy to estimate snowmelt and flow through snow.
- The all-wave radiation module uses shortwave radiation to calculate the net all-wave radiation for input to the evaporation module for snowfree conditions (Granger and Gray, 1990).
- The infiltration module has two parts including Gray's parametric snowmelt infiltration algorithm which estimates snowmelt infiltration into frozen soils (Zhao and Gray, 1999) and Ayers' infiltration that uses soil texture and ground cover as to estimate rainfall infiltration into unfrozen soils (Ayers, 1959).
- The evaporation module uses Granger's evaporation expression to estimate actual evapotranspiration from unsaturated surfaces using an extension of Penman's equation to unsaturated conditions and an energy balance (Granger and Gray 1990 and Granger and Pomeroy, 1997). Evaporation from saturated surfaces is

estimated using the Priestley and Taylor evaporation expression (Priestley and Taylor, 1972).

- The hill slope module uses physically based parameters and principles on hillslopes to calculate subsurface flow and simulate groundwater-surface-water interactions. It is used to calculate soil moisture balance, groundwater storage, subsurface and groundwater discharge, depressional storage and runoff for control volumes of two soil layers, a groundwater layer, and surface depressions (Fang et al, 2010).
- The routing module uses the Muskingum method (Chow, 1964) to route runoff between HRUs in the subbasins using the average flow velocity and average distance from the HRU to the main channel, HRU elevation, overland flow depth and HRU roughness.

The CRHM was used for both the historic and current periods. The Annandale method (Shook and Pomeroy, 2011) was used in place of measured incoming shortwave radiation due to lack of radiation data in the historical period. The model assumes that the land cover is constant and unchanging throughout the modelled period.

3.5 Analysis

3.5.1 Statistical Tests

Annual values of modelled results were examined for the existence of significant trends on a hydrological year time step (Oct-Sep). The missing data from 1987-2005 posed a challenge and so the Mann-Kendall (MK) trend test was chosen for its strength when dealing with non-normally distributed missing data (Hirsch and Slack, 1984; Yue and Pilon 2004). Mann Kendall is a rank-based non-parametric test (Mann, 1945; Kendall, 1975) commonly used to detect monotonic trends in hydrological data (Yue et al., 2002) described as:

$$\tau = \frac{S}{D} \tag{3.7}$$

where

$$S = \sum_{i<j} (sign(x[j] - x[i]) * sign(y[j] - y[i])) \tag{3.8}$$

and

$$D = n\frac{(n-1)}{2} \quad (3.9)$$

where x_j and y_j are the sequential data values and n is the length of the data set. The null hypothesis is that the data is independent and identically distributed while the alternative hypothesis is that a monotonic trend exists.

Due to the longevity of the missing annual values, the difference of mean significance test Mann Whitney U (WU) was chosen as a supplementary test to identify significant changes over time that may have been missed by MK due to the gap in data. Mann Whitney U is a non-parametric test that determines if one distribution is stochastically greater than the other by comparing two means that come from the same population (Mann and Whitney, 1947):

$$U = n_1 n_2 + \frac{n_2(n_2 + 1)}{2} - \sum_{i = n_2 + 1}^{n_2} R_i \quad (3.10)$$

where n_1 is sample size 1, n_2 is sample size 2 and R_i is the rank of the sample size. All p values for both the trend and difference of mean tests less than 0.05 were considered significant.

Autocorrelation can increase the probability of MK finding significant trends. The correction for autocorrelation involves pre-whitening the data set, especially when large gaps are present (Yue et al., 2002). Pre-whitening is used to remove any serial correlation from a data set. To check for the influence of autocorrelation on the trend tests, two data sets were used, one that was pre whitened and one that was not. The influence of autocorrelation proved to be minimal with little to no effect on the results. Because of this it was determined that the non-prewhitened data set would be satisfactory for trend testing. Calculations were performed in the statistical program R using the Kendall (McLeod, 2011) and Stats packages (R Core Team, 2012).

3.5.2 Streamflow

Streamflow was modelled from 2005 to 2013 for comparison to Marmot Creek discharge observations gathered by the Water Survey of Canada. The model results were then analyzed using the Nash-Sutcliffe efficiency (NSE) (Nash and Sutcliffe, 1970), root mean square difference (RMSD), normalized RMSD (NRMSD) and modelled bias (MB) and calculated using the following equations:

$$\text{RMSD} = \sqrt{\frac{1}{n}\sum(X_s - X_o)^2} \tag{3.11}$$

$$\text{NRMSD} = \frac{RMSD}{\bar{X}_o} \tag{3.12}$$

$$\text{MB} = \frac{\sum X_s}{\sum X_o} - 1 \tag{3.13}$$

$$\text{NSE} = 1 - \frac{\sum(X_o - X_s)^2}{\sum(X_o - \bar{X}_o)^2} \tag{3.14}$$

where n are the number of samples and X_o, X_s, $\overline{X_o}$ and $\overline{X_s}$ represent the observed, simulated, mean of the observed, and mean of the simulated values respectively (Fang et al, 2013).

The RMSD is a weighted measure of difference between values, in this case observed versus simulated. The NRMSD is the RMSD that is normalised against the mean of the observed values. A positive MB value indicates over prediction and vice versa. The NSE value is on a scale from 0, estimated values are not different from the average of observed values, to 1, perfect model prediction. This means that any positive NSE value is an indication that the model can predict streamflow, gaining accuracy the closer the value gets to 1. (Fang et al, 2013).

3.5.3 Teleconnections

Correlation analyses were performed between the monthly Pacific Decadal Oscillation (PDO) and El Niño Southern Oscillation (ENSO) values (SOI indices) and the modelled annual values of air temperature, precipitation and the basin hydrological processes using the Pearson coefficient and tested for significance at the 0.05 confidence limit. The Pearson coefficient measures the linear correlation between two variables on a scale from -1, inversely correlated, to 1, positively correlated and 0 meaning no correlation. The PDO index was obtained from http://research.jisao.washington.edu/pdo/PDO.latest, while the SOI was obtained from http://www.esrl.noaa.gov/psd/data/correlation/soi.data. The PDO and ENSO monthly values were broken into two categories; hydrological year (October 1st to September 30th) and winter year (November to March). The Mann Whitney U difference of mean significance test, as outlined above, was also performed on the modelled annual values using a confidence limit of

0.05 when comparing PDO negative regime years and PDO positive regime years to examine the effect of a PDO regime shift on modeled values.

Furthermore, a Generalized Least Square test (GLS) was performed to make regressions of trend, PDO and ENSO with the modelled values (Brockwell and Davis, 2002). The GLS test measures time series regression with serially correlated residuals through the use of predictors (Brockwell and Davis, 2002), which for this study are trend, PDO and ENSO. The test was run sequentially each time taking out a predictor and checking to see if the model performance had changed significantly, if it had then the removed predictor was identified as having a significant effect on the variable being tested against. The point of this test is to attribute changes seen in basin hydrology to various combinations of teleconnections and temporal trend.

CHAPTER 4

RESULTS

This section will describe the results of the statistical tests outlined in Chapter 3, while the interpretation of the results will be discussed in detail in Chapter 5. The hydrometeorological values outlined in this chapter were all simulated using observational inputs in the CRHM as detailed in Chapter 3, with the exception of observed streamflow in section 4.1. The analysis of trends and resiliency were based on an annual basis. Variables such as summer evapotranspiration and winter blowing snow are modeled during their appropriate seasonal months but analyzed annually.

4.1 Streamflow

Streamflow was modelled from 2005 to 2013 for comparison with observed data gathered from Water Survey Canada for Marmot Creek as shown in Figure 4.1. The model results were

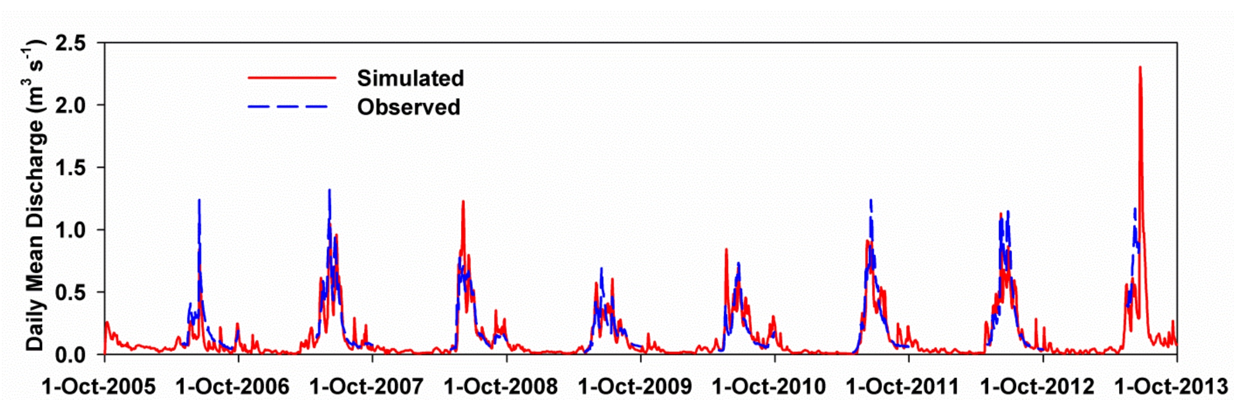

Figure 4.1: Observed and Simulated streamflow for the Marmot Creek Research Basin. The y axis shows the daily mean discharge in m^3/s and the x axis shows the date ranging from 2005 to 2013. (Fang et al., 2016).

then analyzed using the Nash Sutcliffe Error (NSE), the Root Mean Square Deviation (RMSD), normalized RMSD (NRMSD), and modelled bias (MB) to determine their validity (Fang and Pomeroy, 2016). The model does an adequate job of simulating streamflow as shown in the results of Table 4.1. The NSE for all seasons is 0.71 with the lowest being in 2010 at 0.50 and the highest in 2007 at 0.77. The MB values range from -0.39 in 2006 to 0.11 in 2008, but the all-season average is -0.03, which indicates there is no meaningful bias in the model. The RMSD and NRMSD both show small values in comparison with the daily mean discharge from Figure 4.1 which would indicate good model performance. The model had poor performance with extreme years such as 2013 however observational uncertainty was extremely high in the flood conditions of this year.

Table 4.1: Analysis of the simulated streamflow for Marmot Creek including the Nash Sutcliffe Efficiency (NSE), Root Mean Square Difference (RMSD), Normal Root Mean Square Difference (NRMSD) and the Mean Bias (MB) (Fang et al., 2016).

Year	NSE	RMSD	NRMSD	MB
2006	0.63	0.117	0.60	-0.39
2007	0.77	0.141	0.47	-0.09
2008	0.63	0.134	0.50	0.11
2009	0.61	0.093	0.47	-0.01
2010	0.50	0.131	0.64	0.22
2011	0.77	0.136	0.48	-0.02
2012	0.75	0.164	0.52	-0.08
All seasons	0.71	0.133	0.52	-0.03

4.2 Statistical Tests

The MK test was used as the identifier of significant trends. The WU test was used to identify a significant difference of mean between the historic and current periods. The WU test helps strengthen the trends found in the MK test by identifying a significant change between periods. It also displays any possible weakness in the MK test by identifying a significant change in variables where MK did not detect a significant trend.

Table 4.2: Summary of statistical analysis of modelled parameters for trend detection (Mann Kendall (MK)) and difference of mean (Mann-Whitney U (WU)) by HRU. Significantly increasing trends and a significant difference of mean are identified by a X, where yellow means one test detected a significant trend or change and green means both tests detected a significant trend or change.

	Basin		Alpine		Treeline		Upper Forest		Forest Clearing		Lower Forest	
	MK	WU	MK	WU	MK	WU	MK	WU	MK	WU	MK	WU
Air Temperature	X	X	X	X	X	X	X	X	X		X	X
Precipitation	X	X	X	X	X	X		X			X	X
Snowfall	X	X	X	X	X	X						X
Rainfall	X	X	X	X	X	X					X	X
Rainfall Ratio				X		X						
Evapotranspiration	X	X	X	X		X	X	X			X	X
Snow Water Equivalent					X	X						
Blowing Snow Transport					X	X	N/A		N/A		N/A	
Sublimation	X		X		X		N/A		N/A		N/A	
Canopy Sublimation			N/A						N/A			
Snowmelt		X		X	X	X		X			X	X
Runoff		X	X	X	X	X				X		X

4.2.1 Air Temperature

Observed air temperature and precipitation in MCRB have changed over time and indicate a strong significant rising trend (Harder et al., 2015). Changes in modelled air temperature in MCRB between 1969 and 1987 showed a significant a warming trend. Figure 4.2 shows the air temperature for each HRU group and the associated trend line identifies a significant increase. Air temperature has increased at the basin scale and at all elevations with a greater increase at higher elevations. Figure 4.2 shows an increase of almost 2 ^{o}C for every HRU group.

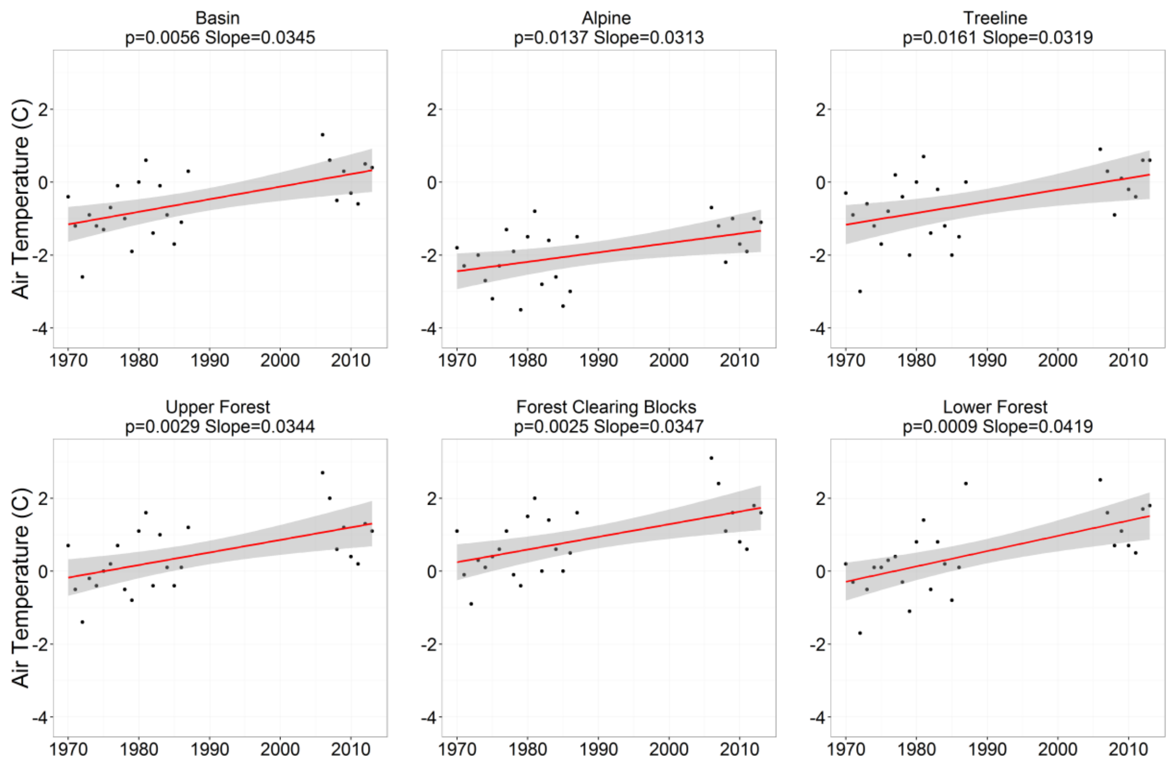

Figure 4.2: Annual air temperature following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale. Red line represents a significant linear trend for Mann Kendall when p< 0.05 and the grey band represents the 95% confidence band for the regression line.

4.2.2 Precipitation

A significant increase in total annual precipitation has also occurred at the basin scale, as well as in the Alpine, Treeline and Lower Forest HRU groups. A significant increase between periods has also been identified for the Alpine, Treeline and Lower Forest HRU groups. Forest Clearing Blocks have shown little to no change throughout the 40 year period. Upper Forest has a p value of 0.06 putting it outside of the null hypothesis for MK but it does show a significant change between periods for WU. This indicates that there was an increase in the Upper Forest but not a significant trend.

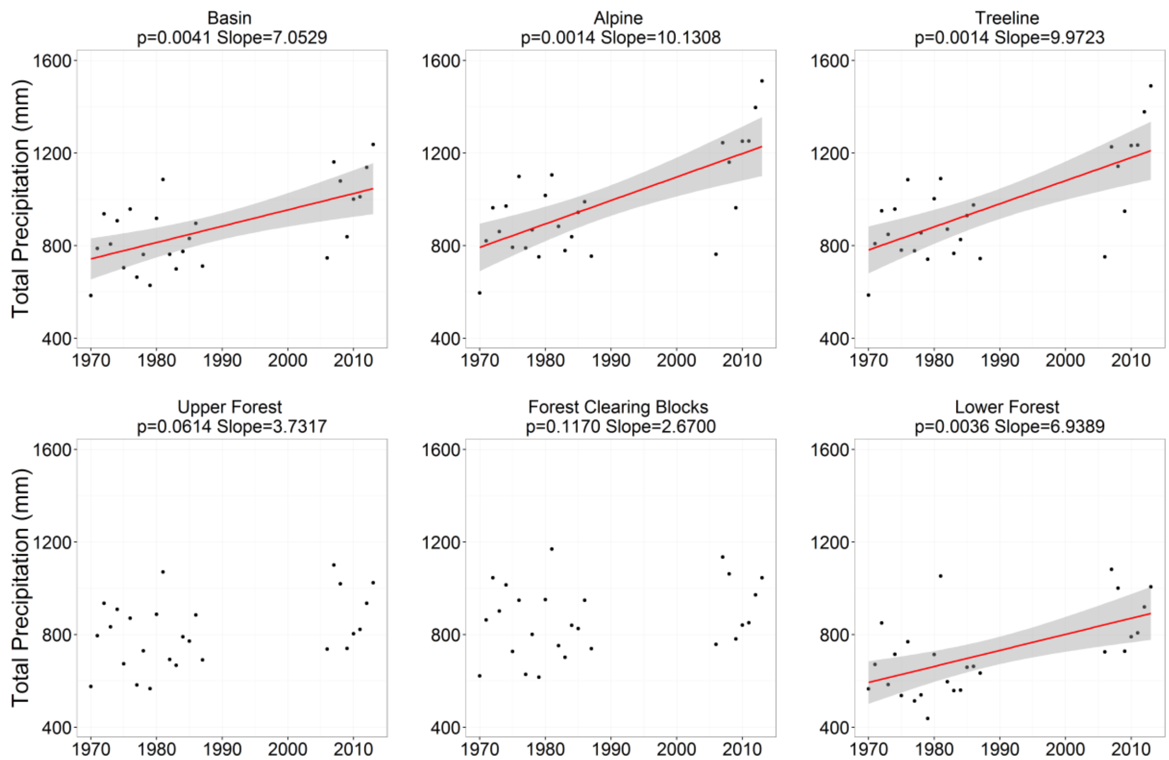

Figure 4.3: Total annual precipitation following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale. Red line represents a significant linear trend for Mann Kendall when $p < 0.05$ and the grey band represents the 95% confidence band for the regression line.

Total annual snowfall, shown in Figure 4.4, is consistent with the precipitation increases by HRU group for the higher elevation sites shown in the Figure 4.3. The higher elevation sites showed a significant trend, while the mid and lower elevation sites did not. The Lower Forest HRU group did not have a statistically significant increase according to MK, but it did show a significant change between periods with the WU test. Total annual rainfall, shown in Figure 4.5, also had a significant trend for both the higher and lower elevation sites. These data indicate that the basin has not only warmed, but became wetter with greater snowfall and rainfall at higher elevations and greater rainfall at lower elevations. Despite the warming temperature and asymmetrical changes in snowfall and rainfall, the percentage of total precipitation that is rainfall (rainfall ratio) did not increase as shown in Figure 4.6. The Treeline HRU group tested positive for a significant change between periods in the WU test.

Figure 4.4: Total annual snowfall following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale. Red line represents a significant linear trend for Mann Kendall when $p < 0.05$ and the grey band represents the 95% confidence band for the regression line.

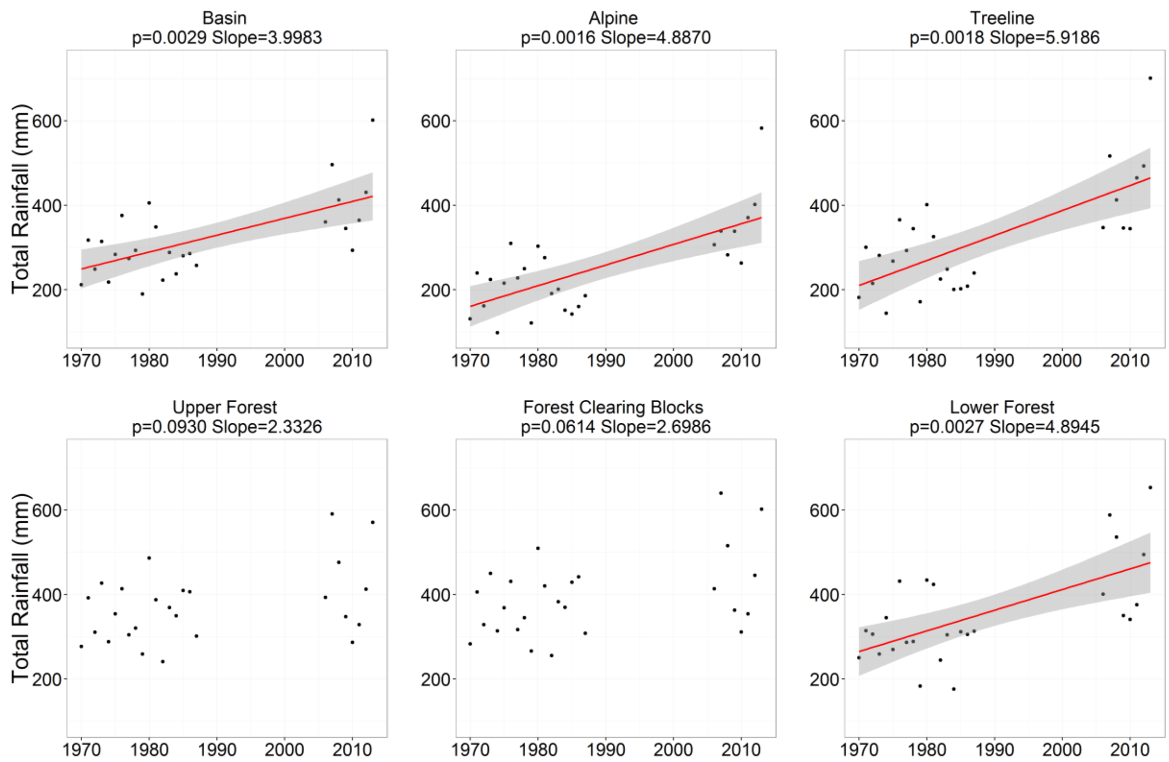

Figure 4.5: Total annual rainfall following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale. Red line represents a significant linear trend for Mann Kendall when $p < 0.05$ and the grey band represents the 95% confidence band for the regression line.

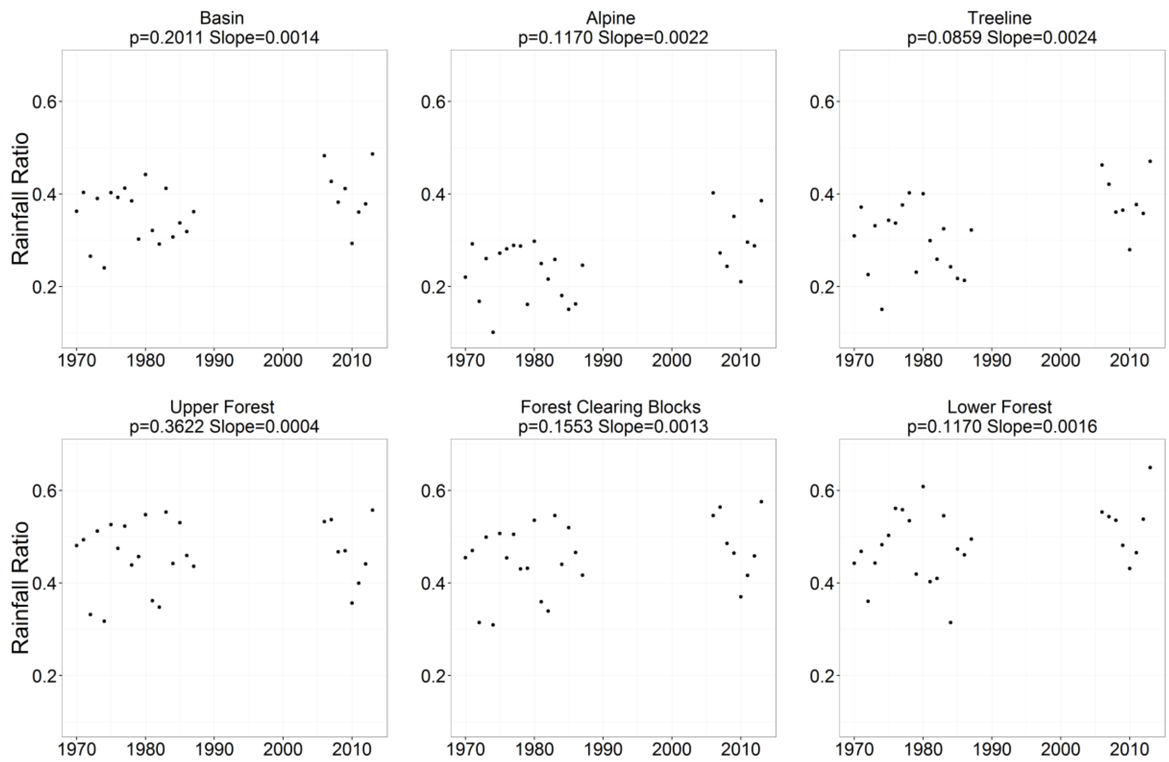

Figure 4.6: Annual rainfall ratio following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale.

4.2.3 Evapotranspiration

Actual evapotranspiration occurring in the summer months underwent a significant increase across the entire basin at every elevation. Figure 4.7 shows an increase in the upper elevation HRU groups Alpine and Upper Forest, as well as, lower in elevation at the Lower Forest. Treeline did not show a significant MK trend, but a significant change between periods did appear in the WU test. Evapotranspiration from the Forest Clearings has generally stayed the same despite an increase in the forested areas surrounding them.

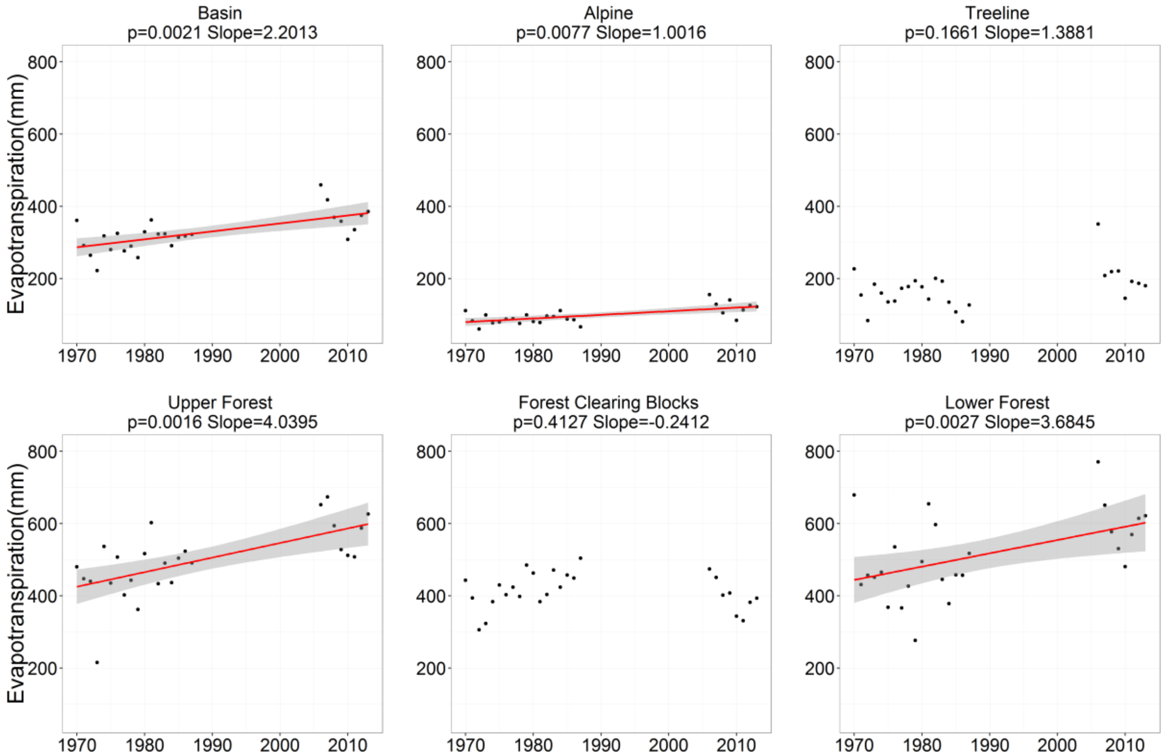

Figure 4.7: Summer evapotranspiration following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale. The red line represents a significant linear trend for Mann Kendall when $p < 0.05$ and the grey band represents the 95% confidence band for the regression line.

4.2.4 Blowing Snow Transport and Sublimation

The MCRB had nearly an equal amount of blowing snow both entering and leaving the basin. However as Figure 4.8 shows, there was a significant increase in the amount of snow blowing into the Treeline resulting in a net loss from the Alpine region. This showed a large amount of redistributed snow collecting in the Treeline, most of which was coming from the Alpine. The Alpine HRU group did not a p value below the 0.05 threshold which means that it lost a large but not significantly changing amount of snow. Figure 4.9 shows that as this snow blew into the Treeline it also sublimated at a significantly increasing rate not only in the Alpine but also at a Basin scale. Only the MK test showed significance for the Basin and Alpine HRU groups.

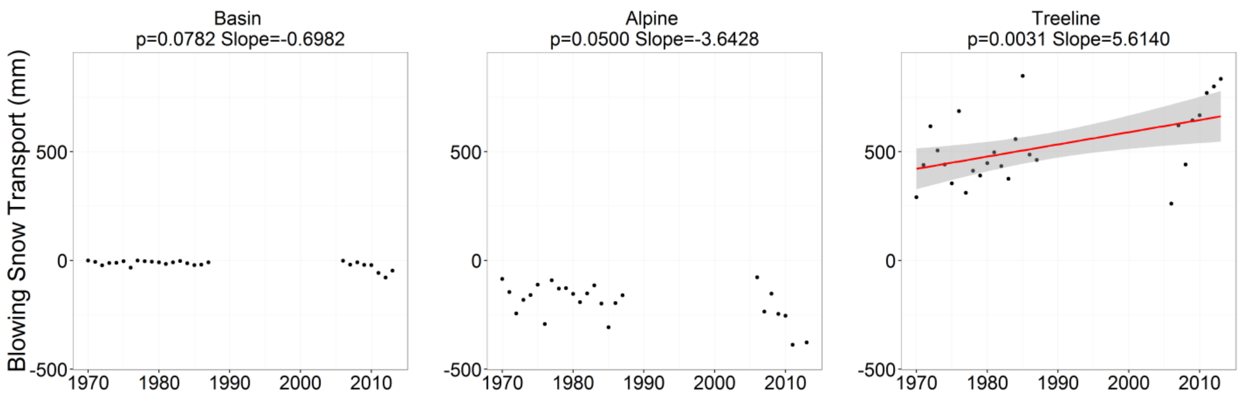

Figure 4.8: Annual blowing snow transport following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale. Upper Forest, Forest Clearing Blocks, and Lower Forest are omitted due to insufficient wind speed to generate blowing snow transport in these HRUs. Red line represents a significant linear trend for Mann Kendall when $p < 0.05$ and the grey band represents the 95% confidence band for the regression line.

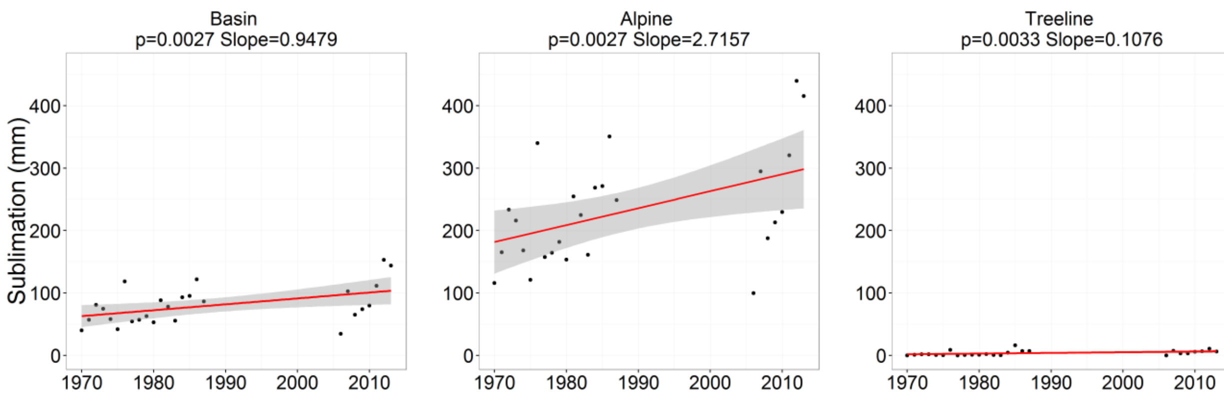

Figure 4.9: Annual sublimation following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale. Upper Forest, Forest Clearing Blocks, and Lower Forest are omitted due to insufficient wind speed to generate blowing snow sublimation in these HRUs. Red line represents a significant linear trend for Mann Kendall when $p < 0.05$ and the grey band represents the 95% confidence band for the regression line.

4.2.5 Intercepted Snow Sublimation

Despite the fact that there was a large amount of snow collecting in the Treeline, the loss to intercepted snow sublimation did not increase – this was partly due to the dominance of deciduous larch at the treeline which intercepts relatively little snow compared to spruce, fir and pine conifer species. This could have also be an indicator that canopy sublimation as a proportion of snowfall was dropping due to warmer winters and more unloading from the canopy during mid-winter thaws. Both the Basin and individual HRU groups all showed substantial variability in their annual values, but none displayed a trend as shown in Figure 4.10.

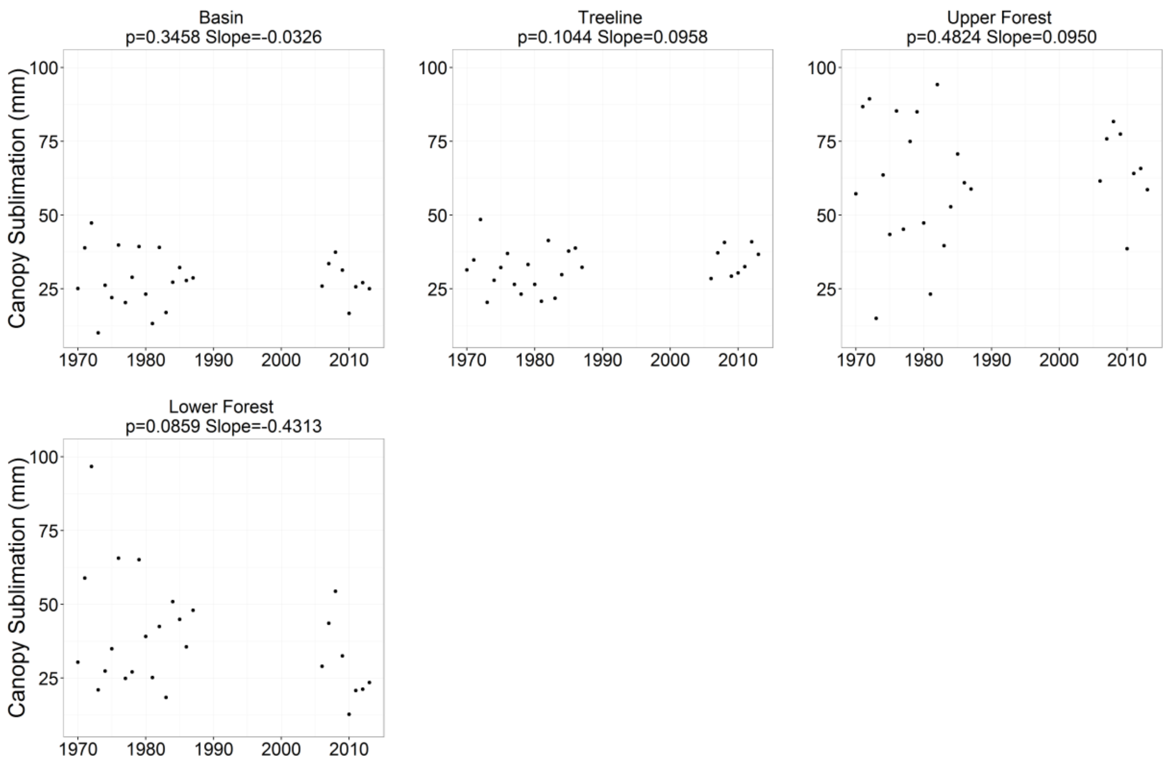

Figure 4.10: Annual canopy sublimation following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale. Alpine and Forest Clearing Blocks omitted due to lack of canopy.

4.2.6 Peak Snow Water Equivalent

Peak SWE showed a significant increase at the Treeline in Figure 4.11where increased blowing snow was collected from the Alpine. There was also a significant change between periods detected at the Treeline by the WU test. Lower Forest and the basin scale both showed an increase, but not significant with p values outside of the 0.05 threshold.

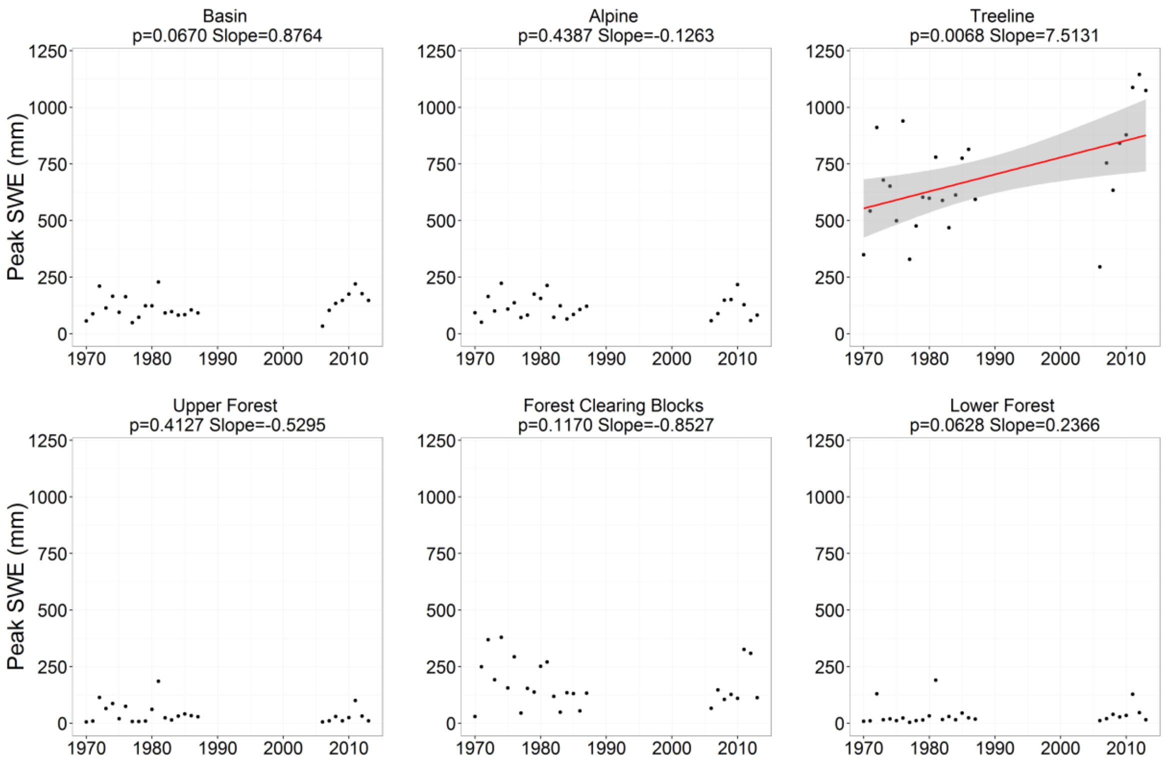

Figure 4.11: Annual peak snow water equivalent following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale. Red line represents a significant linear trend for Mann Kendall when $p < 0.05$ and the grey band represents the 95% confidence band for the regression line.

4.2.7 Snowmelt Volume

Snowmelt volume showed a significant increase in the Treeline and Lower Forest with MK p values well below the 0.05 threshold as shown in Figure 4.12. The rest of the HRU groups, including at the basin scale, appeared to be increasing as well but not at a significant rate according to the MK test. Snowmelt showed a significant change between periods at the Basin scale and in the Upper Forest and Alpine HRU groups according to the WU tests.

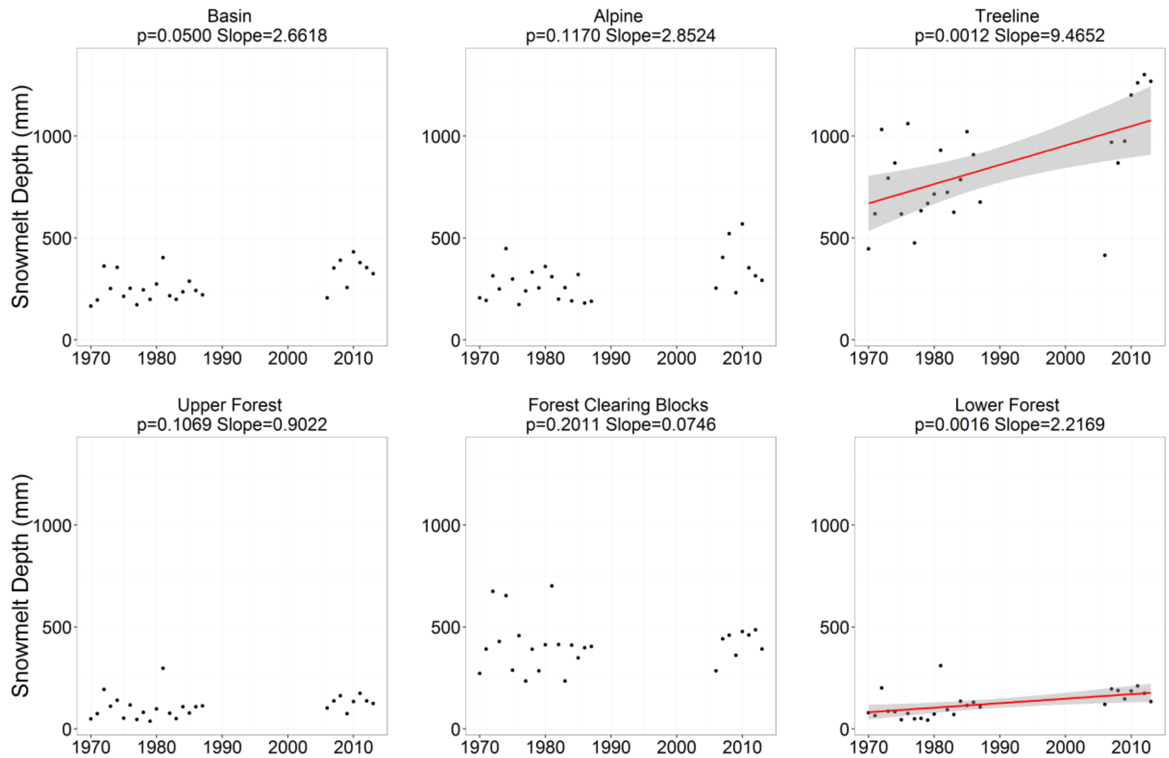

Figure 4.12: Annual snowmelt following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale. Red line represents a significant linear trend for Mann Kendall when $p < 0.05$ and the grey band represents the 95% confidence band for the regression line.

4.2.8 Runoff

Runoff increased significantly in the higher elevations with a p value below the MK 0.05 threshold in the Alpine and Treeline. The WU test identified a significant change between periods in the Lower Forest and Forest Clearing Block HRU groups or at the Basin scale, but Mann Kendall did not find a significant trend. Runoff increased at the basin scale but not significantly. The increase in runoff at the Alpine could be due to the high volume of snow in the higher elevations and rainfall at the lower elevations, with snow being more effective at generating runoff than rainfall, which will then generate more evapotranspiration.

Figure 4.13: Annual runoff following a hydrological year from October 1st to September 30th from Marmot Creek Research Basin (1969 to 2013) grouped by HRU and at the basin scale. Red line represents a significant linear trend for Mann Kendall when $p < 0.05$ and the grey band represents the 95% confidence band for the regression line.

4.3 Water Budget Components

Further assessment of the change in the water budget components was carried out to examine the amount of change over the total study period. Figure 4.14 shows the average amount of each component for both the historic and current period. For their respective input or output, there was an increase in the average of each component from historic to current with the exception of evapotranspiration in the Forest Clearing Blocks. This suggests an overall intensification of the hydrological cycle, including increased precipitation, runoff and associated components, as predicted for climate change (Huntington, 2006). Figure 4.14 gives context for the changes in the water budget components that are displayed in Figure 4.15.

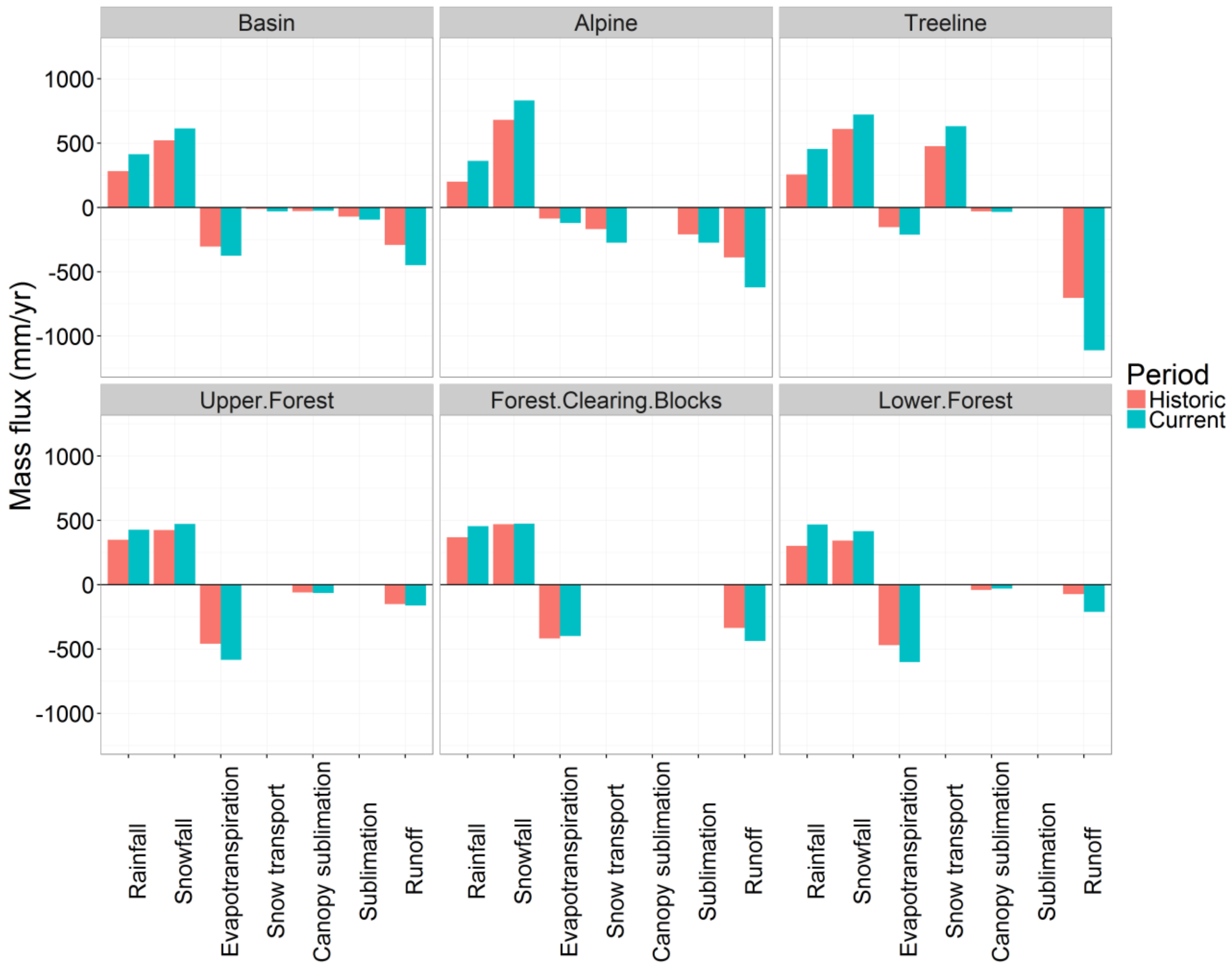

Figure 4.14: Average water balance component for both the historic (1969 to 1987) and current (2005 to 2013) periods broken down by HRU group and at the basin scale.

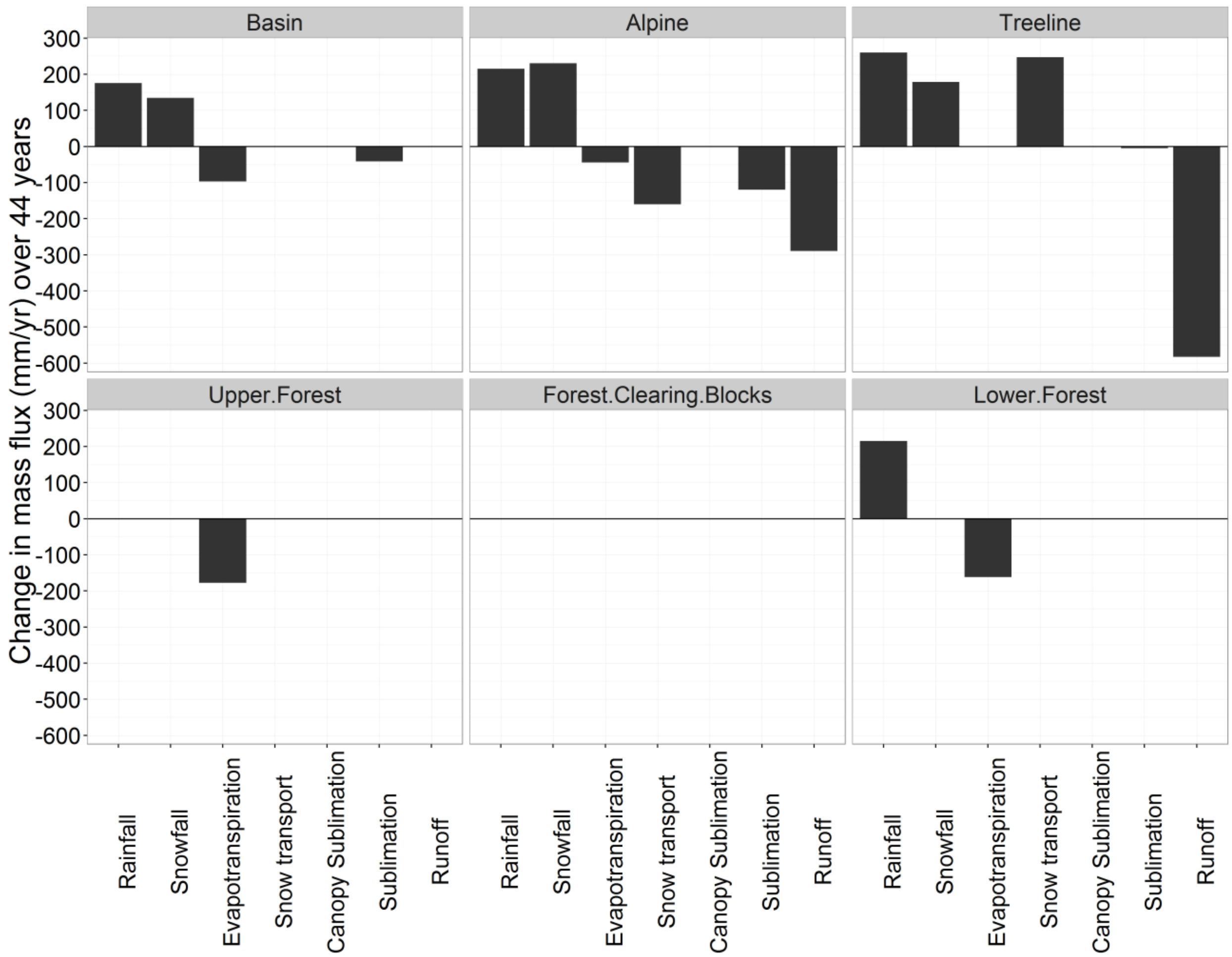

Figure 4.15: Change over the 44 year study period (1969 to 2013) of the water balance components broken down by HRU group and at the basin scale. Only changes that are related to significant trends are shown.

Alpine had a small change in evapotranspiration but large changes in precipitation which were counterbalanced by large changes in runoff, sublimation and snow blowing into the Treeline. Treeline had a large change in rainfall, runoff, incoming snow transport and snowfall which was counterbalanced by a significant change in evapotranspiration in the Upper Forest and Lower Forest. Lower Forest also displayed a significant change in rainfall. Forest Clearing Blocks do not have a significant trend in any water balance component. This is consistent with the trend at the basin scale which showed a significant change in precipitation that was being counterbalanced by evapotranspiration in the spring and summer and sublimation during the winter months.

4.4 Teleconnections

Correlation and regression analysis was performed between annual modelled hydrometeorological variables and water balance component values and the PDO and ENSO on both hydrological year (Oct to Sep) and winter (Nov to Mar) periods. For the PDO, as shown in Table 4.3, on an annual basis precipitation at the higher elevations showed a significant inverse correlation, but not at the lower elevations or the overall basin. At the high elevations, snow water equivalent, canopy sublimation, snowmelt and runoff all showed a significant inverse correlation with the PDO while blowing snow transport showed a significant positive correlation. However, in the lower elevations PDO was only significantly correlated with evapotranspiration and inversely correlated with snow water equivalent, specifically in the Forest Clearing Blocks.

Table 4.4 shows that the significant correlation between PDO winter values (November to March) and hydrometeorological variables differs slightly than on an annual basis. Total precipitation still showed an inverse significant correlation with the PDO, but at a basin scale and not in the higher elevations. Canopy sublimation was not significant at the Treeline, but was inversely correlated in the Upper Forest and at the basin scale. Runoff was still significant at the Treeline with an inverse correlation, but not at in the Alpine or at the basin scale. Evapotranspiration showed a significant positive correlation in the Forest Clearing Blocks. Snow water equivalent still showed a significant inverse correlation in the Forest Clearing Blocks as well as snow water equivalent at all elevations and the basin scale.

As shown in Table 4.5, on an annual basis the SOI showed a significant inverse correlation with blowing snow transport at the basin scale and a significant positive correlation with snow water equivalent in the forest clearing blocks and canopy sublimation at all elevations and on a basin scale. Compared to the annual PDO significance, the ENSO did not show a strong relationship with hydrological processes and water balance components in the basin. The ENSO winter values (November to March), in Table 4.6, are very similar to the annual values with a significant positive correlation with snow water equivalent in the Forest Clearing Blocks and with canopy sublimation at all elevations and the basin scale.

Table 4.3 – Correlation between annual modelled values versus PDO indices values on a hydrological year (Oct to Sep) time step tested at the 0.05 confidence limit. P value is displayed with the r correlation value in brackets.

	Basin	Alpine	Treeline	Upper Forest	Forest Clearing Blocks	Lower Forest
Air Temperature	0.38(0.18)	0.54(0.12)	0.51(0.13)	0.26(0.23)	0.30(0.21)	0.34(0.19)
Total Precipitation	0.05(-0.39)	**0.04(-0.41)**	**0.04(-0.41)**	0.12(-0.31)	0.13(-0.31)	0.07(-0.36)
Snowfall	0.09(-0.34)	0.10(-0.33)	0.12(-0.31)	0.12(-0.31)	0.16(-0.28)	0.18(-0.27)
Rainfall	0.12(-0.31)	0.06(-0.37)	0.07(-0.36)	0.47(-0.15)	0.43(-0.16)	0.11(-0.32)
Rainfall Ratio	0.77(-0.06)	0.30(-0.21)	0.36(-0.19)	0.46(0.15)	0.60(0.11)	0.68(-0.08)
Evapotranspiration	0.81(-0.05)	0.25(-0.24)	0.45(-0.15)	0.77(-0.06)	**$4.1*10^{-4}$ (0.64)**	0.50(-0.14)
Snow Water Equivalent	**0.04(-0.41)**	0.96(-0.01)	**0.03(-0.42)**	0.76(-0.06)	**0.01(-0.5)**	0.63(-0.10)
Blowing Snow	**0.02(0.45)**	**0.04(0.40)**	0.12(-0.32)	N/A	N/A	N/A
Sublimation	0.48(-0.15)	0.47(-0.15)	0.99(0.00)	N/A	N/A	N/A
Canopy Sublimation	0.08(-0.35)	N/A	**0.04(-0.41)**	0.05(-0.39)	N/A	0.39(-0.18)
Snowmelt	0.07(-0.36)	0.15(-0.29)	**0.04(-0.41)**	0.43(-0.16)	0.15(-0.29)	0.51(-0.14)
Runoff	**0.04(-0.40)**	**0.04(-0.41)**	**0.03(-0.42)**	**0.47(-0.15)**	**0.06(-0.37)**	**0.26(-0.23)**

Table 4.4 – Correlation between annual modelled values versus PDO indices values on a winter period (Nov to Mar) time step tested at the 0.05 confidence limit. P value is displayed with the r correlation value in brackets.

	Basin	Alpine	Treeline	Upper Forest	Forest Clearing Blocks	Lower Forest
Air Temperature	0.31(0.21)	0.41(0.17)	0.38(0.18)	0.24(0.24)	0.27(0.22)	0.32(0.2)
Total Precipitation	**0.04(-0.4)**	0.05(-0.39)	0.05(-0.39)	0.08(-0.35)	0.06(-0.37)	0.07(-0.36)
Snowfall	0.06(-0.38)	0.09(-0.34)	0.08(-0.35)	0.06(-0.37)	0.06(-0.38)	0.16(-0.28)
Rainfall	0.16(-0.28)	0.11(-0.32)	0.16(-0.29)	0.50(-0.14)	0.47(-0.15)	0.10(-0.33)
Rainfall Ratio	0.99(0.00)	0.49(-0.14)	0.70(-0.08)	0.33(0.2)	0.40(0.17)	0.64(-0.1)
Evapotranspiration	0.88(-0.03)	0.48(-0.15)	0.70(-0.08)	0.66(-0.09)	**$3.3*10^{-3}$ (0.55)**	0.73(-0.07)
Snow Water Equivalent	**$4.9*10^{-3}$ (-0.53)**	0.37(-0.18)	**0.01(-0.48)**	0.35(-0.19)	**$3.6*10^{-4}$ (-0.65)**	0.39(-0.18)
Blowing Snow	**0.02(0.44)**	**0.04(0.41)**	0.10(-0.33)	N/A	N/A	N/A
Sublimation	0.38(-0.18)	0.37(-0.18)	0.98(0)	N/A	N/A	N/A
Canopy Sublimation	**0.02(-0.44)**	N/A	0.05(-0.39)	**0.02(-0.45)**	N/A	0.13(-0.3)
Snowmelt	0.08(-0.35)	0.30(-0.21)	0.03(-0.43)	0.31(-0.21)	0.05(-0.38)	0.54(-0.13)
Runoff	0.10(-0.33)	0.13(-0.31)	**0.04(-0.4)**	0.51(-0.14)	0.07(-0.36)	0.38(-0.18)

Table 4.5 – Correlation between annual modelled values versus SOI indices values on a hydrological year (Oct to Sep) time step tested at the 0.05 confidence limit. P value is displayed with the r correlation value in brackets.

	Basin	Alpine	Treeline	Upper Forest	Forest Clearing Blocks	Lower Forest
Air Temperature	0.51(-0.13)	0.48(-0.14)	0.44(-0.16)	0.47(-0.15)	0.50(-0.14)	0.73(-0.07)
Total Precipitation	0.15(0.29)	0.16(0.28)	0.16(0.28)	0.21(0.26)	0.17(0.27)	0.16(0.29)
Snowfall	0.08(0.35)	0.16(0.28)	0.10(0.33)	0.06(0.37)	0.06(0.38)	0.14(0.29)
Rainfall	0.58(0.12)	0.42(0.17)	0.55(0.12)	0.93(-0.02)	0.98(-0.01)	0.34(0.2)
Rainfall Ratio	0.48(-0.14)	0.92(0.02)	0.75(-0.07)	0.12(-0.32)	0.15(-0.29)	0.79(-0.06)
Evapotranspiration	0.52(0.13)	0.42(0.17)	0.91(0.02)	0.37(0.18)	0.07(-0.36)	0.53(0.13)
Snow Water Equivalent	0.05(0.39)	0.73(0.07)	0.06(0.37)	0.17(0.28)	**$2.2*10^{-3}$ (0.57)**	0.31(0.21)
Blowing Snow	**0.04(-0.40)**	0.07(-0.36)	0.17(0.28)	N/A	N/A	N/A
Sublimation	0.30(0.21)	0.30(0.21)	0.49(0.14)	N/A	N/A	N/A
Canopy Sublimation	**0.02(0.45)**	N/A	**0.03(0.43)**	**0.02(0.47)**	N/A	0.14(0.29)
Snowmelt	0.18(0.27)	0.79(0.05)	0.10(0.33)	0.18(0.27)	0.06(0.37)	0.33(0.20)
Runoff	0.43(0.16)	0.67(0.09)	0.22(0.25)	0.77(0.06)	0.20(0.26)	0.53(0.13)

Table 4.6 – Correlation between annual modelled values versus SOI indices values on a winter period (Nov to Mar) time step tested at the 0.05 confidence limit. P value is displayed with the r correlation value in brackets.

	Basin	**Alpine**	**Treeline**	**Upper Forest**	**Forest Clearing Blocks**	**Lower Forest**
Air Temperature	0.43(-0.16)	0.43(-0.16)	0.41(-0.17)	0.39(-0.17)	0.43(-0.16)	0.60(-0.11)
Total Precipitation	0.13(0.31)	0.14(0.30)	0.14(0.30)	0.18(0.27)	0.16(0.28)	0.16(0.28)
Snowfall	0.08(0.35)	0.13(0.30)	0.10(0.33)	0.07(0.36)	0.08(0.35)	0.19(0.27)
Rainfall	0.50(0.14)	0.40(0.17)	0.48(0.14)	0.92(0.02)	0.85(0.04)	0.29(0.22)
Rainfall Ratio	0.53(-0.13)	0.96(0.01)	0.80(-0.05)	0.16(-0.28)	0.22(-0.25)	0.89(-0.03)
Evapotranspiration	0.47(0.15)	0.31(0.21)	0.72(0.07)	0.37(0.18)	0.06(-0.37)	0.45(0.16)
Snow Water Equivalent	0.09(0.34)	0.87(0.03)	0.08(0.35)	0.29(0.22)	**$4.6*10^{-3}$ (0.54)**	0.51(0.13)
Blowing Snow	0.05(-0.39)	0.08(-0.35)	0.19(0.27)	N/A	N/A	N/A
Sublimation	0.34(0.20)	0.34(0.20)	0.52(0.13)	N/A	N/A	N/A
Canopy Sublimation	**0.02(0.45)**	N/A	**0.03(0.44)**	**0.01(0.48)**	N/A	0.22(0.25)
Snowmelt	0.19(0.27)	0.62(0.10)	0.11(0.32)	0.27(0.23)	0.09(0.34)	0.44(0.16)
Runoff	0.33(0.20)	0.53(0.13)	0.20(0.26)	0.65(0.09)	0.16(0.29)	0.45(0.15)

PDO regime shifts were then analyzed for accompanying changes in the modelled annual values using the Mann Whitney U (WU) significance test with significance being identified with a p value < 0.05 (Tables 4.5 and 4.6). Evapotranspiration in the Forest Clearing Blocks was the only variable found to have a significant increase between regime shift years during this test. These test results indicate that when the PDO has a regime shift from negative to positive or positive to negative it causes a significant increase in evapotranspiration in this specific HRU. This was only found in the WU test and was not consistent with findings in the other teleconnections analysis.

Table 4.7 – Mann Whitney U PDO regime change significance test ($p < 0.05$) of annual modelled values and their associated PDO positive (1976 to 87) and negative (1970 to 76 and 2005 to 13) years Highlighted years are increasing significantly.

	Basin	**Alpine**	**Treeline**	**Upper Forest**	**Forest Clearing Blocks**	**Lower Forest**
Air Temperature	0.78	0.80	0.78	0.73	0.77	0.79
Total Precipitation	0.96	0.92	0.92	0.95	0.93	0.99
Snowfall	0.86	0.88	0.84	0.92	0.81	0.99
Rainfall	0.95	0.96	0.96	0.77	0.78	0.96
Rainfall Ratio	0.77	0.90	0.93	0.55	0.70	0.57
Evapotranspiration	0.91	0.96	0.97	0.95	**0.01**	0.96
Snow Water Equivalent	0.92	0.52	0.89	0.36	0.78	0.55
Blowing Snow	0.19	0.23	0.77	N/A	N/A	N/A
Sublimation	0.41	0.41	0.46	N/A	N/A	N/A
Canopy Sublimation	0.33	N/A	0.79	0.63	N/A	0.15
Snowmelt	0.93	0.97	0.86	0.95	0.88	0.96
Runoff	1.00	1.00	0.95	0.99	1.00	1.00

The Generalized Least Square test was used to test regressions of trend, PDO and ENSO with the modelled and measured values. Trend was assumed to be due to gradual climate change whereas PDO and ENSO were due to cyclical changes via teleconnections with oceanic influences on local weather. The results in Table 4.7 indicate that at the basin scale the only significant term was trend while PDO and ENSO showed no consistent basin-wide impact on any hydrological process or on basin scale runoff.

Table 4.8 – Generalized Least Squares test identifying the statistical significant predictors (ENSO, PDO or Trend) of the MCRB modelled values. Only the HRUs for each variable that tested positive for one of the three terms (PDO, ENSO or Trend) are shown.

Variable	HRU Group	Significant Terms
Air Temperature	Basin	Trend
	Alpine	Trend
	Treeline	Trend
	Upper Forest	ENSO, Trend
	Forest Clearing	Trend
	Lower Forest	Trend
Total Precipitation	Basin	Trend
	Alpine	Trend
	Treeline	Trend
	Upper Forest	Trend
	Lower Forest	Trend
Snowfall	Basin	Trend
	Alpine	Trend
	Treeline	Trend
Rainfall	Basin	ENSO, Trend
	Alpine	Trend
	Treeline	Trend
	Upper Forest	Trend
	Lower Forest	Trend
Rainfall Ratio	Alpine	Trend
	Treeline	Trend
	Forest Clearing	Trend
Evapotranspiration	Basin	Trend
	Alpine	Trend
	Upper Forest	Trend
	Lower Forest	Trend
Snow Water Equivalent	Upper Forest	Trend
Blowing Snow Transport	Treeline	ENSO, PDO, Trend
Canopy Sublimation	Treeline	PDO, Trend
Snowmelt	Basin	Trend
	Alpine	PDO, Trend
	Treeline	Trend
	Lower Forest	Trend
Runoff	Basin	Trend
	Alpine	Trend
	Treeline	ENSO, Trend
	Lower Forest	Trend

CHAPTER 5
DISCUSSION

Marmot Creek Research Basin has a dramatically changing climate that has had an important effect on some hydrological processes. A rising air temperature paired with an increase in precipitation has facilitated changes in specific hydrological processes. The importance of these changes is that they are occurring at different elevations and compensate for one another leading to little effect on the overall basin runoff.

Air temperature was found to have increased across the basin, and at all elevations. The increase in temperature coincides with previous research inside the MCRB (Harder et al., 2015) and also outside in other regions of the Rocky Mountains (St. Jacques et al., 2009; Woo and Pomeroy, 2011 The greatest increase in air temperature has occurred in the forested and forest clearing areas which are located in the lower elevation of the basin, this coincides with Harder et al. (2015). A significant increase in total annual precipitation was also evident at the basin scale, as well as in the Alpine, Treeline and Lower Forest HRU groups for all three trend tests. Forest Clearing Blocks showed little to no change in precipitation.. Upper Forest showed an increase as well but it was insignificant as the MK p value fell just outside of the threshold. However, there are differences in the amount of precipitation found at different elevations in the basin when compared with a similar study (Harder et al., 2015). It is clear that despite the lack of evidence in the forest clearings, the basin became wetter across all elevations.

Snowfall and rainfall both increased at the basin scale and in the Alpine and Treeline regions. At lower elevations rainfall significantly increased but snowfall only showed a significant change between periods that was evident in the Mann Whitney-U test. Despite observations of increasing air temperature, the rainfall ratio did not increase significantly at all elevations which differs from what other studies have found in the region (Shook and Pomeroy, 2012) and across the Canadian Prairies (Dumanski et al., 2015). The Alpine and Treeline HRU groups did show a significant change between periods in the WU test.

Actual evapotranspiration can only occur once the snowfree period starts, and showed a significant increase across the entire basin at every elevation. This was to be expected with an increase in air temperature across the basin. However, Forest Clearing Blocks did not show an increase and generally had low variability over the whole time period despite an increase in evapotranspiration in the forest surrounding them. This may be due to the limited vegetation in clearings and its inability to transpire rapidly, even if climatic conditions permit higher evapotranspiration rates (Fang et al., 2013).

There was a proportionate amount of blowing snow both entering and leaving the basin with very low variability from year to year. There was a significant increase over time in the snow redistributed to the Treeline which coincides with a significant decrease in blowing snow transport in the Alpine. However, these trends are not equivalent as there was a significant increase in Alpine sublimation which limits increases in the snow available for redistribution to the Treeline. There was uncertainty in the historical wind speed reconstruction and the blowing snow process, which varies with wind speed to the fourth power (Pomeroy et al., 1993), is especially sensitive. Appropriate instrumentation in the current period allows for more certain estimates of blowing snow. However, comparing the historic blowing snow to the current period blowing snow shows plausible results. At the Treeline, blowing snow transport showed a significant increase over time in the MK test and a significant change between periods in the WU test. Intercepted snow sublimation showed a large variability in annual values in the forested regions but no significant increases or significant changes between periods. Sublimation in the forest is complicated by the unloading process and this can lead to a nonlinear response to rising air temperature (Hedstrom and Pomeroy, 1998).

Peak snow water equivalent only increased significantly at the Treeline as a result of the increased blowing snow transport. This was not evident in the observations from Harder et al. (2015) but the difference may be because the snow surveys were not specific to the treeline or issues with the data collection in the past (Oltmann, 1997). In the Lower Forest snowmelt increased while peak SWE did not, which contradicts Harder et al. (2015) as peak SWE observations showed a very significant decrease. It is possible this is due to forest regrowth where the early snow surveys were collected, but the cause remains unknown and the uncertainty of historical snow survey measurements is high. In contrast to the observations, the model did not capture the decrease in peak SWE but it did show a temperature response as total melt was increasing. This is likely due to warmer air temperatures increasing the prevalence of midwinter melts which results in minimal changes in snow accumulation and an increase in total snowmelt (Mote et al., 2005). This might not be inconsistent with declining measured SWE peaks, particularly when these are weekly or monthly in some cases and ablation has occurred before the snow survey.

Runoff showed a significant increase at high elevations as well as a significant change between periods at all elevations and at the Basin scale. Much of the runoff production in MCRB is snowmelt-dominated so the increase in the Alpine runoff suggests a changing regime that favours more rainfall-runoff. The increase in total precipitation at the Alpine offsets the loss of blowing snow to the Treeline which implies a greater rainfall runoff response in the upper elevation.

The resiliency of basin streamflow in the study period is due to counter balancing among the significantly increasing water balance components. At the basin scale precipitation increased significantly over time, as indicated by the trend tests, but evapotranspiration and sublimation were also significantly increasing and serving as a counterbalance. At a HRU level, Alpine and Treeline showed the largest amount of interaction among water balance components. An increase of over 400 mm of total precipitation was observed in Alpine over the study period, but this gain was counterbalanced by sublimation, evapotranspiration and a net loss of blowing snow transport into the Treeline. The Treeline also had a significant increase in precipitation as rainfall and snowfall increased over time and snow blew in from the Alpine totalling almost 600 mm. However, this accumulated precipitation at the Treeline was counterbalanced by a significant increase in evapotranspiration in the forested areas. The total size of the Upper Forest and Lower Forest HRUs means that even a small change in evapotranspiration in the forested areas could account for such a large change in runoff over a smaller area such as Treeline. The Forest Clearing Blocks seemed to be unaffected by anything and Upper Forest had a net loss to evapotranspiration. Lower Forest showed a significant increase of over 200 mm in rainfall which was counterbalanced by evapotranspiration.

Teleconnections displayed a significant correlation with a portion of the hydrological processes in the MCRB. Correlations were found between the PDO and the SOI for various hydrometeorological variables in the hydrological year (October to September) and winter year (November to March) periods. A correlation test examined the response of meteorological variables, air temperature and precipitation, which drive the modelled changes in the hydrological processes. The PDO had a positive correlation with air temperature and an inverse correlation with precipitation which is expected as positive PDO values should result in warmer and dryer conditions (Bonsal et al., 2001). The SOI (ENSO) had an inverse correlation with air temperature and a positive correlation with precipitation which is also expected as positive SOI values should result in cooler and wetter conditions (Bonsal et al., 2001).

On a hydrological year time step, the PDO showed a significant correlation with precipitation in the higher elevations which affected other significantly correlated processes such as snow water equivalent, blowing snow drift, snowmelt, canopy sublimation and runoff. At the basin scale snow water equivalent, blowing snow transport and runoff were also significantly correlated as well as evapotranspiration in the Forest Clearing Blocks. The type of correlation was consistent for the hydrological variables. An inverse correlation with snow water equivalent, snowmelt, canopy sublimation and runoff is consistent with the inverse correlation with precipitation, while a positive correlation with evapotranspiration is consistent with a positive correlation with air temperature. Blowing snow transport is largely dependent on wind and will not have a clear relationship with air temperature and precipitation. On a winter time step (Nov-Mar), the PDO showed similar significant correlations, and the same relationships, to the annual time step with a few changes; total precipitation and snow water equivalent were significant at the basin scale and not at a specific elevation, canopy sublimation changed from being significantly correlated at the Treeline to the Upper Forest and at the basin scale, runoff was only significantly correlated at the Treeline and snowmelt was no longer significantly correlated at the Treeline.

The SOI, on a hydrological year time step, was significantly correlated with blowing snow transport and canopy sublimation at the basin scale, as well as canopy sublimation in the Treeline and Upper Forest, and evapotranspiration in the Forest Clearing Blocks. On a winter time step (Nov-Mar), the SOI was significantly correlated with the same variables as the hydrological year time step except blowing snow transport at the basin scale. A positive correlation between the SOI and evapotranspiration and canopy sublimation is plausible as a warmer temperature would increase evaporation and sublimation rates. This differed from the sublimation conditions with the PDO since precipitation had an inverse relationship with the PDO whereas with the SOI it has a positive correlation, meaning a positive SOI value with wetter conditions and more snow to sublimate. The Pearson coefficient test on hydrological variables produced significant correlations in the Marmot Creek Research Basin. However, these significant correlations, while plausible, were only shown to affect a small portion of the compensatory processes behind the water balance and had no effect on streamflow generation. There are studies that relate changes in air temperature and precipitation in Western Canada to large scale teleconnections (Bonsal et al., 2001; Bonsal and Shabbar, 2004) however, these relationships are more difficult to relate to small scale hydrological processes in a mountain basin. For a more suitable comparison large scale hydrological responses at the river basin scale would need to be compared to the large scale PDO and ENSO.

The Mann Whitney-U PDO regime shift test showed a statistically significant change between regime shift years and evapotranspiration in the Forest Clearing Blocks. The GLS test has been shown to identify the significance of trend, PDO and ENSO (St. Jacques, 2010; Harder et al. 2015) terms on a variable of interest. Trend was indicated as a significant predictor for the majority of the variables in agreement with the MK trend test and WU significance test. ENSO was shown as a predictor for Treeline drift and runoff, Upper Forest air temperature and Basin rainfall. Harder et al. (2015) identified ENSO as a predictor for low elevation precipitation and PDO as a predictor for peak SWE, neither of which appeared as significant in this study. The weak correlation between teleconnections and annual hydrometeorological variables, the limited significance of the GLS test results, and inconsistency with Harder et al. (2015), indicates that teleconnections have a weak connection with basin hydrology.

Increases in precipitation have been compensated for by increases in evapotranspiration. This explains a mechanism behind the resilience of the hydrology of Marmot Creek that has been also noted by Harder et al. (2015) using basin observations, but no diagnoses that might explain the resilience. Changes in climate have affected hydrological processes and produced significant trends at all elevations and land cover types. However, these trends have worked to counterbalance each other resulting in a basin that has seen significant change, while streamflow generation has been consistent over time. For a snowmelt driven basin this streamflow resiliency, is of great importance for the water security of downstream users (Wheater and Gober, 2013).

CHAPTER 6

CONCLUSIONS

The impact of climate change on basin hydrology in the MCRB has been minimized by the resiliency of the basin through the counterbalancing of the water budget components, specifically sublimation and evapotranspiration. Annual meteorological and hydrological process values were analyzed at varying elevations and grouped according to terrain type. Hydrological processes showed great fluctuations from year to year, many with an increasing trend. Meteorological variables were also showing a changing climate with an increasing trend in air temperature and precipitation. Despite all of this change, runoff from the basin remained mostly unaffected, displaying great resilience to climate change.

The average temperature has been rising significantly over time and has had an effect on the hydrological processes of the basin. There have also been changes in the basin hydrology based on altitude. Higher and lower elevation sites have yielded different results for individual processes. There was a significant increase in the depth of precipitation at the basin scale at all elevations. Rainfall fraction of precipitation did not increase at any elevation or at the basin scale. Peak snow accumulation decreased in the Alpine and increased at the Treeline due to greater wind redistribution, but had no basin-wide trend. Blowing snow transport increased from the Alpine to the Treeline with concomitant sublimation losses. Runoff did not increase from the basin, but increased significantly from the Alpine and Treeline.

Snowmelt volume increased across the basin and was significantly increased specifically at the Treeline and Lower Forest. Evapotranspiration increased basin-wide and at all elevations with the greatest increase at low elevations. There is evidence for compensatory processes leading to greater snow accumulation, melt and runoff particularly at high elevations and greater evapotranspiration loss at low elevations with greater sublimation loss at high elevations.

Water budget components were analyzed and compared from the historic and current period as well as the significant increase in these components. The average of each water balance component has increased from the historic to the current period, with the exception of runoff in the Forest Clearing Blocks. The significant increase over time and counterbalance of these components was largest in the Alpine-Treeline interaction. Significant increases of over 400 mm in precipitation were counterbalanced by significant increases in sublimation, blowing snow transport and evapotranspiration. This is also evident in the lower elevation forests where rainfall and evapotranspiration significantly increase at a similar rate to counterbalance each other. The water budget components expose these counter balancing significant trends and show the cause behind the resiliency of the basin streamflow to climate change.

There was a significant correlation between teleconnections and the modeled hydrometeorological variables in the MCRB, however, these significant correlations were only shown to affect a small portion of the compensatory processes behind the water balance. Specifically the affected variables were precipitation, evapotranspiration, snow water equivalent, blowing snow transport, canopy sublimation and runoff for PDO and snow water equivalent, blowing snow transport and canopy sublimation for ENSO at varying elevations. The positive and inverse relationships between the PDO and SOI indices with these water balance components showed that the cause of some of the change in basin hydrology may be linked to teleconnections. This correlation was not evident however in the GLS test where very few variables were shown to be significantly influenced by PDO and ENSO. The WU regime shift test also failed to detect significant changes in any variables besides evapotranspiration in the Forest Clearing Blocks. The lack of consistency between statistical tests combined with only a small amount of the total variables showing a correlation could be attributed to scaling issues when comparing large scale teleconnections with small scale basin hydrology. However, based on the findings of this study the conclusion is that teleconnections have no meaningful impact on basin hydrology and the processes responsible for generating streamflow.

From this analysis it is evident that there are a large number of contributing factors to changes in basin hydrology. Rising temperatures and increased precipitation both are driving factors behind the changes to the hydrological processes in Marmot Creek. The results show that the increase in precipitation in the basin is compensated by increased evapotranspiration, sublimation and snow transport. And while rising temperature has an impact on and changes the hydrological processes within the basin, the counterbalance of water budget components results in a streamflow that shows resiliency to climate change. This study canbe used as a stepping stone for further exploration into other basins of similar size and geography to see if the Marmot Creek Research Basin is the standard or anomaly for climate change.

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Appendix A Teleconnections Tables

Table A.1 – Coefficient of determination (r^2) value for Hydrological Year (Oct-Sep) annual modelled values versus PDO indices value.

	Basin	Alpine	Treeline	Upper Forest	Forest Clearing Blocks	Lower Forest
Air Temperature	0.03	0.01	0.02	0.05	0.05	0.04
Total Precipitation	0.20	0.17	0.17	0.10	0.10	0.13
Snowfall	0.12	0.11	0.10	0.10	0.08	0.07
Rainfall	0.10	0.14	0.13	0.02	0.03	0.10
Rainfall Ratio	0.00	0.04	0.03	0.02	0.01	0.00
Evapotranspiration	0.00	0.06	0.02	0.00	0.41	0.02
Snow Water Equivalent	0.17	0.00	0.18	0.00	0.26	0.01
Blowing Snow	0.21	0.16	0.10	0.00	0.00	0.00
Sublimation	0.02	0.02	0.00	0.00	0.00	0.00
Canopy Sublimation	0.13	0.00	0.17	0.20	0.00	0.03
Snowmelt	0.13	0.08	0.17	0.03	0.08	0.02
Runoff	0.16	0.17	0.18	0.02	0.14	0.05

Table A.2 – Coefficient of determination (r^2) value for Winter Year (Nov-Mar) annual modelled values versus PDO indices value.

	Basin	Alpine	Treeline	Upper Forest	Forest Clearing Blocks	Lower Forest
Air Temperature	0.04	0.03	0.03	0.06	0.05	0.04
Total Precipitation	0.16	0.15	0.15	0.12	0.14	0.13
Snowfall	0.14	0.11	0.12	0.14	0.14	0.08
Rainfall	0.08	0.10	0.08	0.02	0.02	0.11
Rainfall Ratio	0.00	0.02	0.01	0.04	0.03	0.01
Evapotranspiration	0.00	0.02	0.01	0.01	0.31	0.01
Snow Water Equivalent	0.29	0.03	0.23	0.04	0.42	0.03
Blowing Snow	0.20	0.17	0.11	0.00	0.00	0.00
Sublimation	0.03	0.03	0.00	0.00	0.00	0.00
Canopy Sublimation	0.19	0.00	0.15	0.20	0.00	0.09
Snowmelt	0.12	0.04	0.19	0.04	0.15	0.02
Runoff	0.11	0.09	0.16	0.02	0.13	0.03

Table A.3 – Coefficient of determination (r^2) value for Hydrological Year (Oct-Sep) annual modelled values versus SOI indices value.

	Basin	Alpine	Treeline	Upper Forest	Forest Clearing Blocks	Lower Forest
Air Temperature	0.02	0.02	0.02	0.02	0.02	0.01
Total Precipitation	0.09	0.08	0.08	0.07	0.08	0.08
Snowfall	0.12	0.08	0.11	0.14	0.14	0.09
Rainfall	0.01	0.03	0.01	0.00	0.00	0.04
Rainfall Ratio	0.02	0.00	0.00	0.10	0.08	0.00
Evapotranspiration	0.02	0.03	0.00	0.03	0.13	0.02
Snow Water Equivalent	0.15	0.01	0.14	0.08	0.33	0.04
Blowing Snow	0.16	0.13	0.08	N/A	N/A	N/A
Sublimation	0.04	0.04	0.02	N/A	N/A	N/A
Canopy Sublimation	0.21	N/A	0.18	0.22	N/A	0.09
Snowmelt	0.07	0.00	0.11	0.07	0.14	0.04
Runoff	0.03	0.01	0.06	0.00	0.07	0.02

Table A.4 – Coefficient of determination (r^2) value for Winter Year (Nov-Mar) annual modelled values versus SOI indices value.

	Basin	Alpine	Treeline	Upper Forest	Forest Clearing Blocks	Lower Forest
Air Temperature	0.03	0.03	0.03	0.03	0.03	0.01
Total Precipitation	0.09	0.09	0.09	0.07	0.08	0.08
Snowfall	0.12	0.09	0.11	0.13	0.12	0.07
Rainfall	0.02	0.03	0.02	0.00	0.00	0.05
Rainfall Ratio	0.02	0.00	0.00	0.08	0.06	0.00
Evapotranspiration	0.02	0.04	0.01	0.03	0.14	0.02
Snow Water Equivalent	0.11	0.00	0.12	0.05	0.29	0.02
Blowing Snow	0.15	0.12	0.07	N/A	N/A	N/A
Sublimation	0.04	0.04	0.02	N/A	N/A	N/A
Canopy Sublimation	0.20	N/A	0.19	0.23	N/A	0.06
Snowmelt	0.07	0.01	0.10	0.05	0.12	0.02
Runoff	0.04	0.02	0.07	0.01	0.08	0.02

Appendix B Regression Equations

Table B.1: Regression equations for the historical sites (Twin1, Cab5 and Con5) and their associated r^2 values. Where at and AT are equal to air temperature, u is equal to wind speed and vp is equal to vapour pressure.

Site	**Equations**	**r^2**
Confluence 5 (Con5)	at_Con5 = 0.993023*AT_Cab5 + 1.26153	0.87
	at_Con5 = 1.00793*AT_Twin1 + 1.4416	0.81
	u_Con5 = 0.104288*u_Twin1 + 0.940853	0.26
	u_Con5 = 0.510456*u_Cab5 + 0.514064	0.43
	vp_Con5 = 0.849063*vp_Cab5 + 0.0937889	0.64
	vp_Con5 = 0.976901*vp_Twin1 + 0.100734	0.61
Cabin 5 (Cab5)	at_Cab5 = 0.87503*at_Con5 - 1.27693	0.87
	at_Cab5 = 0.973349*at_Twin1 + 0.128994	0.85
	u_Cab5 = 0.153376*u_Twin1 + 1.09871	0.32
	u_Cab5 = 0.835479*u_Con5 + 0.534439	0.43
	vp_Cab5 = 0.751567*vp_Con5 + 0.0940237	0.64
	vp_Cab5 = 0.945348*vp_Twin1 + 0.100734	0.61
Twin 1 (Twin1)	at_Twin1 = 0.875619*at_Cab5 - 0.309927	0.85
	at_Twin1 = 0.802171*at_Con5 - 1.44342	0.81
	u_Twin1 = 2.46937*u_Con5 + 0.662247	0.26
	u_Twin1 = 2.11478*u_Cab5 + 0.312983	0.32
	vp_Twin1 = 0.643639*Cab5 + 0.0865269	0.61
	vp_Twin1 = 0.605914*Con5 + 0.0875622	0.60