An evaluation of winter hydroclimatic variables conducive to snowmelt and the generation of extreme hydrologic events in western Canada

by

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M.Sc., University of Victoria, 2014
B.Sc., University of Alberta, 2011

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Supervisory Committee

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Abstract

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The frequency, magnitude, and atmospheric drivers of winter hydroclimatic conditions conducive to snowmelt in western Canada were evaluated. These hydroclimatic variables were linked to the mid-winter break-up of river ice that included the creation of a comprehensive database including 46 mid-winter river ice break-up events in western Canada (1950-2008) and six events in Alaska (1950-2014). Widespread increases in above-freezing temperatures and spatially diverse increases in rainfall were detected over the study period (1946-2012), particularly during January and March. Critical elevation zones representing the greatest rate of change were identified for major river basins. Specifically, low-elevation (500-1000 m) temperature changes dominated the Stikine, Nass, Skeena, and Fraser river basins and low to mid-elevation changes (700-1500 m) dominated the Peace, Athabasca, Saskatchewan, and Columbia river basins. The greatest increases in rainfall were seen below 700 m and between 1200-1900 m in the Fraser and at mid- to high-elevations (1500-2200 m) in the Peace, Athabasca, and Saskatchewan river basins. Daily synoptic-scale atmospheric circulation patterns were classified using Self-Organizing Maps (SOM) and corresponding hydroclimatic variables were evaluated. Frequency, persistence, and preferred shifts of identified synoptic types provided additional insight into characteristics of dominant atmospheric circulation patterns. Trend analyses revealed significant ($p < 0.05$) decreases in two dominant synoptic types: a ridge of high pressure over the Pacific Ocean and adjacent trough of low pressure over western Canada, which directs the movement of cold, dry air over the study region, and zonal flow with westerly flow from the Pacific Ocean over the study region. Conversely, trend analyses revealed an increase in the frequency and persistence of a ridge of high pressure over western Canada over the study period. However, step-change analysis revealed a decrease in zonal flows and an increase in the occurrence of high-pressure ridges over western Canada in 1977, coinciding with a shift to a positive Pacific Decadal Oscillation regime. A ridge of high pressure over western Canada was associated with a high frequency and magnitude of above-freezing temperatures and rainfall in the study region. This pattern is highly persistent and elicits a strong surface climate response. A ridge of high pressure and associated above-freezing temperatures and rainfall was also found to be the primary driver of mid-winter river ice break-up with rainfall being a stronger driver west of the Rocky Mountains and temperature to the east. These results improve our understanding of the drivers of threats to snowpack integrity and the generation of extreme hydrologic events.

Table of Contents

List of Tables

List of Figures

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Chapter 1: Introduction

1.1. Background

Ice and snow are critical components of the hydrologic cycle in cold regions. The radiative and thermal properties of ice and snow have a profound influence on terrestrial and aquatic environments (Walsh *et al.* 2005; Wrona *et al.* 2005; Callaghan *et al.* 2011a; Prowse *et al.* 2011a), and are involved in critical feedback mechanisms that influence local to global climate (Callaghan *et al.* 2011b; Prowse *et al.* 2011b; Derksen *et al.* 2012; Cohen *et al.* 2012). Snow acts as an insulator, protecting terrestrial vegetation and habitat from harsh winter temperatures (Walsh *et al.* 2005; Callaghan *et al.* 2011a; Vincent *et al.* 2011), while freshwater ice dynamics are essential to aquatic ecosystems, affecting water temperature, primary productivity, and biogeochemical cycles (Wrona *et al.* 2005; Lind *et al.* 2014; Prowse *et al.* 2011a). Moreover, many regions, including western Canada, are dependent upon seasonal frozen water storage and release to provide water resources to meet municipal, industrial, and agricultural needs (Barnett *et al.* 2005; Stewart 2009).

Concern has been raised over changes to temperature and precipitation patterns in western Canada (Prowse *et al.* 2013; Sauchyn and Kulshreshtha 2008; Walker and Sydneysmith 2008), particularly given the disproportionately high rate of change during the winter season (Zhang *et al.* 2000; Vincent *et al.* 2015; O'Neil *et al.* 2017). There have been decreases in the number of frost days ($T_{min} < 0°C$) and snowfall fraction and increases in the number of days with rain (Vincent and Mekis 2006; Zhang *et al.* 2011; Mekis and Vincent 2011; Vincent *et al.* 2015). O'Neil *et al.* (2017) detected statistically significant decreases in snow accumulation and melt in southwestern Canada, coinciding with earlier spring river ice break-up (de Rham *et al.* 2008; von de Wall 2011). Evidence indicates declines in snow cover extent, particularly during spring (Brown and Mote 2009; Hernández-Henríquez *et al.* 2015; Kang *et al.* 2016; Najafi *et al.* 2017), and at low- to mid-elevations in the Western Cordillera (Brown and Mote 2009; Hernández-Henríquez *et al.* 2015). Additionally, there have been increases in the frequency and duration of winter (Jan-Mar) warm spells (Shabbar and Bonsal 2003) and in mid-winter river ice break-up events (Beltaos 2002; Prowse and Beltaos 2002). These trends signify broad-scale threats to water security and the potential for the generation of hydrologic extremes, giving rise to the need to better understand the role of climatic variability on hydrologic systems in western Canada.

Snowmelt represents a nonlinear response to temperature and/or precipitation where an increase in temperature will not elicit a phase change from solid to liquid unless a particular threshold temperature has been exceeded and with sufficient energy inputs to the snowpack. The temperature of the snowpack dominates the energy inputs required to melt snow during the winter, such that a sub-freezing snowpack requires energy to first raise the temperature of the snowpack to 0°C-isothermal before a phase change is initiated (USACE 1956; Male and Gray 1981; DeWalle and Rango 2008). Precipitation phase, solid, liquid, or mixed, determines whether snow accumulates or melts, while the temperatures of both the rain and the snowpack influence the volume of any resulting runoff (DeWalle and Rango 2008). Humidity, with or without precipitation, can contribute to snowmelt through sensible and latent heat energy transfers (USACE 1956; Bruce and Clark 1966; Harpold *et al.* 2017). Understanding the hydroclimatic conditions conducive to winter snowmelt is essential for the management and prediction of available water resources. Additionally, improved knowledge of the generation of extreme hydrologic events during the winter season, including river ice break-up and flooding, is essential for the mitigation of the impacts of such extremes and will aid in determining whether the hydroclimatic triggers of these events have increased in concert with changing climate over recent decades.

Patterns of atmospheric pressure and the rotation of the earth are fundamental forces behind the movement of air masses and water vapour. The mid-troposphere is characterized by a series of mid-latitude troughs and ridges resulting in meridional flow, or, in the absence of troughs and ridges, zonal flow. Troughs and ridges direct the movement of high- and low- pressure systems at the surface and are indicative of the exchange of warm, moist tropical air and cold, dry polar air over the mid-latitudes (Holton 1979). Additionally, mountain ranges are large physical barriers to surface air movement, resulting in orographic uplift and precipitation on the windward side, and dry, adiabatically warmed air on the leeward side (Longley 1967; Goulding 1978), causing instability and lee cyclogenesis (Martin 2006), and this is reflected in the surface climate of western Canada.

Synoptic climatology is a powerful discipline for evaluating the role of atmospheric circulation on surface hydroclimatic conditions and has been instrumental in enhancing our understanding of atmospheric drivers of mountain snowpack (e.g., Chagnon *et al.* 1993; McGregor and Gellatly 1996; Moore and McKendry 1996; Fitzharris *et al.* 1997), glacier mass balance (e.g., Yarnal 1984; Moore and Demuth 2001; Shea and Marshall 2007), avalanches (e.g., Fitzharris 1987; Birkeland *et al.* 2001; Yokely *et al.* 2014), lake-effect snowfall (e.g., Leathers and Ellis 1996; Kunkel *et al.* 2002; Liu and Moore 2004), and surface air temperature and precipitation (e.g., Hewitson and Crane 2002; Hope *et al.* 2006; Finnis *et al.* 2009; Jónsdóttir and Uvo 2009; Cassano and Cassano 2010; Newton *et al.* 2014; Bonsal *et al.* 2017). Synoptic climatology involves either classifying a large number of atmospheric circulation patterns into a smaller set of dominant patterns, then relating those patterns to surface climate (e.g., Newton *et al.* 2014; Bonsal *et al.* 2017), or identifying the surface climate or hydrologic variable of interest and creating a composite of atmospheric circulation corresponding to that variable (e.g., Bonsal *et al.* 2001; Shabbar *et al.* 2011). Both the 'circulation-to-environment and 'environment-to-circulation' approaches provide valuable information about atmosphere-surface relationships, and the method applied depends on the objectives of the research.

Snowmelt and rainfall runoff are the primary drivers behind the break-up of river ice (Gray and Prowse 1993; Beltaos and Prowse 2001; Beltaos 2008; de Rham *et al.* 2008; Beltaos 2013). These driving forces must exceed the resisting forces, including the strength and thickness of the ice and attachment to riverbanks, for break-up to occur (Beltaos 1997; Beltaos and Prowse 2001; Beltaos 2003; Beltaos 2008). The most severe river ice break-up events tend to be driven by high-intensity snowmelt, which typically occur during spring (Prowse and Marsh 1989; Beltaos 2008).

Hydroclimatic conditions generating snowmelt runoff - above-freezing temperatures and/or rain-on-snow - have the potential to trigger river ice break-up during the winter (Prowse *et al.* 1990b; Beltaos and Prowse 2001; Beltaos 2002). The effect of solar radiation is negligible during the winter; therefore, river ice is typically strong and intact, and break-up is mechanical and may lead to the formation of ice jams and flooding

(Beltaos 2002; Beltaos 2003). The mid-winter break-up of river ice has been studied extensively in eastern Canada and the United States (e.g., Beltaos 1999; Beltaos 2002; Huntington *et al.* 2003; Carr and Vuyovich 2014), while research in western Canada has tended to focus on case studies (e.g., Doyle 1988; Doyle 1992; Janowicz 2010). Prowse *et al.* (2002) identified a temperate zone across Canada and United States with an increased risk of mid-winter river ice break-up based on predetermined accumulated freezing degree-day thresholds, and projected a shift in the boundaries of the temperate zone due to climate change. However, a comprehensive analysis of mid-winter river ice break-up events in western Canada remains a gap in the research. Given the strong relationship between air temperature and river ice break-up (Prowse and Beltaos 2002; Bonsal and Prowse 2003), knowledge of the atmospheric drivers of winter surface climate (e.g., Newton *et al.* 2014), particularly snow accumulation (e.g., Romolo *et al.* 2006a) and melt (e.g., Romolo *et al.* 2006b), is fundamental to understanding the hydroclimatic conditions conducive to mid-winter river ice break-up.

1.2. Study area

The headwaters of several major river basins in western Canada are located in the Western Cordillera. The Liard, Peace, and Athabasca rivers are major tributaries to the north-flowing Mackenzie River. The Liard River flows from the Pelly Mountains in southeastern Yukon, through northeastern British Columbia before shifting northeast through the Northwest Territories to join the Mackenzie River. The Peace and Athabasca rivers originate on the eastern slopes of the Rocky Mountains and flow east to the Peace-Athabasca Delta (Peters *et al.* 2006). The Peace River is regulated by the W.A.C. Bennett dam, which alters seasonal streamflow, including lower spring peak flow and higher winter flow (Peters and Prowse 2001). The Saskatchewan River is a tributary to the Nelson River, draining into Hudson Bay. The Saskatchewan River originates on the eastern slopes of the Rocky Mountains and flows over a semi-arid prairie region and is heavily allocated, primarily for agricultural purposes (Martz *et al.* 2007; Sauchyn and Kulshreshtha 2008).

The Fraser, Columbia, Stikine, Nass, and Skeena rivers flow west, draining into the Pacific Ocean. The Fraser River flows northwest in the Rocky Mountain Trench from its headwaters in the Rocky Mountains, before shifting south, flowing through central British Columbia. The central Fraser River is located in a semi-arid grassland, while the upper and lower Fraser receive higher precipitation (Reynoldson *et al.* 2005). The Columbia River originates in southeastern British Columbia and flows south through the United States before draining into the Pacific Ocean. Although a relatively small portion of the Columbia River is located in Canada, a large portion of the annual streamflow originates from this headwater region (Pederson *et al.* 2011). The Stikine, Nass, and Skeena are smaller watersheds located in the north coastal region of British Columbia.

1.3. Dissertation Objectives

Despite numerous studies evaluating hydroclimatic trends and variability in western Canada, few studies have focused on temperature-related thresholds conducive to snowmelt, or the atmospheric drivers of such variables. Furthermore, although several case studies of mid-winter river ice break-up events, including causes and impacts, have been examined, a broad-scale evaluation of these events has not been completed. Due to the infrequent, discrete, and spatially variable nature of extreme hydrologic events such as mid-winter river ice break-up, it is difficult to evaluate trends; however, the trends and variability of the hydroclimatic and atmospheric drivers of these events can be evaluated. Understanding these hydroclimatic variables informs water resource management and the mitigation of extreme hydrologic events. This research is designed to fill critical knowledge gaps by providing a comprehensive analysis of the following:

1) spatial and temporal patterns of snowmelt-related winter hydroclimatic variables, specifically above-freezing temperatures and rainfall, with an emphasis on identifying critical elevation zones representing a higher risk of snowmelt, in major watersheds in western Canada using high-resolution gridded temperature and precipitation data from 1946-2012;
2) dominant daily synoptic-scale atmospheric circulation patterns from 1949-2012, a description of characteristics such as frequency, persistence, trajectory, and temporal trends and changes, and identify patterns of associated frequency and magnitude of surface above-freezing temperatures and rainfall;
3) mid-winter river ice break-up events in western Canada and Alaska from 1950-2008 (western Canada) and 1950-2014 (Alaska) including analysis of hydroclimatic and atmospheric drivers.

1.4. Thesis format

This thesis is divided into five chapters. The first chapter provides an introduction and context for this research. Chapters 2, 3, and 4 are stand-alone articles representing the three phases of this research. Chapter 2 focuses on the spatial and temporal trends and variability in winter hydroclimatic variables conducive to snowmelt. Chapter 3 examines the atmospheric drivers of these hydroclimatic variables. Chapter 4 is focused on identifying and evaluating mid-winter river ice break-up events in western Canada and Alaska, and was published in the Journal Hydrology Research (Newton *et al.* 2017). The thesis concludes with Chapter 5, synthesizing results from Chapters 2, 3, and 4, and recommending future research directions.

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Chapter 2: Spatial and temporal trends in above-freezing winter temperatures and associated rainfall that control snowpack conditions in western Canada

Abstract

Winter snowpack is central to water resource availability in western Canada. Hydroclimatic conditions associated with snowmelt threaten the integrity of the snowpack and can lead to the generation of extreme hydrologic events. This research evaluates trends and variability in winter (December - March) above-freezing temperatures and associated rainfall in western Canada using a daily, high-resolution gridded dataset from 1946-2012. Days when the mean daily temperature is above freezing ($T_{mean} > 0°C$) were identified and accumulated melting degree days (MDD) calculated. Rainfall was evaluated using a temperature-index equation and a threshold temperature of 1°C. Results reveal a significant widespread increase in the frequency of winter days when the mean daily temperature is above freezing and accumulated MDD, particularly during January and March. Significant increases in rainfall are primarily confined to the coast and portions of the Columbia, Fraser and Peace river basins, and are dominated by increases during March. In the north coastal (Stikine, Nass, and Skeena) and Fraser river basins the rate of increase in the frequency and magnitude of above-freezing temperatures are greatest at low elevations (500-1000 m). For rainfall, the greatest increases are below 700 m and between 1200 and 1900 m in the Fraser and below 1000 m in the north coastal basins. The Peace, Athabasca, Saskatchewan and Columbia river basins see the greatest increases in above-freezing temperatures at low to

mid-elevations (700-1500 m) and minor increases in rainfall, primarily at higher elevations (1500-2200 m). This research improves our understanding of threats to snowpack integrity and the potential generation of hydrologic extremes, particularly at these critical elevation zones.

2.1. Introduction

Winter snow accumulation and melt are key components of the hydrologic cycle in cold regions. Climate change has already had profound impacts on snowpack regimes, representing one of the greatest risks to water security (e.g., water supply, floods, low flows) in western Canada (Sauchyn and Kulshreshtha 2008; Walker and Sydneysmith 2008). Biophysical and socio-economic systems are dependent upon water resource availability, the majority of which originates as snowpack, particularly in mountain headwaters (Barnett *et al.* 2005; Stewart 2009). Water resource availability is critical for the generation of hydroelectricity (Filion 2000; Martz *et al.* 2007), agricultural productivity (Pentney and Ohrn 2008), groundwater recharge (Hayashi and van der Kamp 2005) and aquatic ecosystem health (Wrona *et al.* 2006, Burn *et al.* 2008; Wrona *et al.* 2016). Moreover, reductions in snowpack may contribute to the length and severity of summer drought conditions (Bonsal *et al.* 2011; Hanesiak *et al.* 2011). Snow accumulation and melt are sensitive to fluctuations in air temperature, and warm season water shortages resulting from low snowpack or early snowmelt pose challenges for water resource management at a time when water demand is highest (Schindler and Donahue 2006).

There is a growing concern over the frequency, magnitude, and impacts of extreme hydrologic events (Bates *et al.* 2008). The onset and end of the snow accumulation season and river ice freeze- and break-up are strongly linked to temperature shifts above and below 0°C (Prowse and Beltaos 2002; Bonsal and Prowse 2003; Brown and Mote 2009). Air temperatures rising above freezing, freeze-thaw cycles, and rain-on-snow during the cold season, can threaten the integrity of the snowpack. Additionally, these hydroclimatic conditions can lead to sublimation and initiate snowmelt (Gray and Male 1981; Gray and Prowse 1993), triggering hydrologic extremes including river ice break-up (Beltaos 2002; Newton *et al.* 2017) and flooding (Anderson and Larson 1996; Marks *et al.* 1998; Shabbar and Bonsal 2003; McCabe *et al.* 2007), and can lead to landslides (Harr 1981; Christner and Harr 1982) and snow avalanches (Fitzharris 1987; Hägeli and McClung, 2003; Jamieson *et al.* 2001). Winter flooding is particularly hazardous as flood management challenges can be amplified by the presence of snow and ice and the return of sub-freezing temperatures (Beltaos 2002).

Extreme hydroclimatic events are, by definition, infrequent, high-magnitude occurrences (Zhang *et al.* 2011a); however, a cold season hydrologic extreme can occur even when the trigger itself is not extreme, emphasizing the nonlinear nature of system responses. Recent studies of trends in temperature and precipitation extremes have focused on the high and low ends of the distributions at a particular location, such as changes in the frequency of temperature above the 90th or below the 10th percentile and precipitation above the 95th or 99th percentile values (Alexander *et al.* 2006; Vincent and Mekis 2006; Zhang *et al.* 2011a; Donat *et al.* 2013). This definition of extremes describes relative changes and is non-stationary as the temperature and precipitation distributions are dependent on the given time series climatology or reference period. Absolute, or stationary, thresholds describing above/below freezing temperatures and rainfall/snowfall are more meaningful for snow accumulation and ice growth or snow and ice melt, and, consequently, the generation of hydrologic extremes.

Spatial and temporal patterns of temperature and precipitation have changed over recent decades with the greatest rates of change occurring during winter and at higher latitudes (Zhang *et al.* 2000; O'Neil *et al.* 2017). Changing temperature and precipitation patterns have altered the climatological average and the shifts in distribution have had profound impacts on extreme high and low temperature and precipitation (Vincent and Mekis 2006; Zhang *et al.* 2011a; Zhang *et al.* 2011b). Even minor shifts in spatial and temporal patterns of precipitation can have major consequences for water availability at the watershed scale, including changes to the timing and magnitude of high and low streamflow (Prowse *et al.* 2013; Bawden *et al.* 2015; O'Neil *et al.* 2017). Additionally, research has revealed changes to seasonal ice-cover including earlier spring break-up (de Rham *et al.* 2008; von de Wall 2011) and increased frequency of mid-winter river ice break-up events across Canada and the United States (Beltaos 2002; Prowse *et al.* 2002; Carr and Vuyovich 2014).

Recent evidence indicates declining spring snow-cover extent and an earlier spring freshet in major river basins in western Canada. Statistically significant decreases in snow cover extent have been documented in North America (Déry and Brown 2007; Brown and Mote 2009; Hernández-Henríquez *et al.* 2015), particularly in the Western Cordillera (Brown and Mote 2009; Choi *et al.* 2010). Vincent *et al.* (2015) reported both declines in the ratio of snowfall to total precipitation and snow-cover duration in western Canada from 1948-2012, particularly during the spring. In the Fraser River Basin, Kang *et al.* (2016) detected decreasing trends in the magnitude of mountain snowpack and an earlier onset of snowmelt and spring freshet from 1949-2006. Similarly, Kang *et al.* (2014) found that decreasing snow-water equivalent (SWE) resulted in a lower contribution of snowmelt to total runoff in the Fraser River. The onset of snowmelt was significantly earlier in the upper Peace River Basin from 1963-1996, attributed to changes in dominant large-scale atmospheric circulation patterns (Romolo *et al.* 2006). Najafi *et al.* (2017) detected declines in spring (1 April) SWE from 1950-2006 in the upper Peace, Fraser, and upper Columbia basins, with the greatest trends occurring at elevations below 1200 m. Concomitantly, significant earlier trends in the timing of the spring freshet have been found for numerous rivers in western Canada (Whitfield 2001; Zhang *et al.* 2001; Stewart *et al.* 2005; Abdul Aziz and Burn 2006; Burn 2008; Rood *et al.* 2008). The general hydroclimatic trends observed in western Canada have been observed in other mountainous regions including the Himalayas (Azmat *et al.* 2017; Muhuri *et al.* 2017), China (Zhang *et al.* 2017), the Andes (Mernild *et al.* 2017; Vuille *et al.* 2018), and the Alps (Klein *et al.* 2016; Marty *et al.* 2017).

The volume of runoff resulting from a temperature- or rainfall-driven melt events is a function of the depth, density, and temperature of the snowpack. In particular, dominant energy processes are dependent on whether the snowpack is warm (0°C-isothermal) or cold (<0°C; USACE 1956; DeWalle and Rango 2008). During winter months, the temperature of the snowpack is typically sub-freezing and energy from solar radiation is negligible (USACE 1956). Potential sources of heat energy to the snowpack includes convective heat transfer from overlying air, relatively warm rain water, and/or condensation on the snowpack during periods of high humidity (USACE 1956; DeWalle and Rango 2008; Harpold *et al.* 2017). In a 0°C-isothermal snowpack, the energy supplied to the snowpack is used to melt the snow and runoff is generated. Conversely, when the snowpack is cold, sensible heat energy is initially used to warm the snow and while the snowpack remains below freezing, the rainwater freezes within the snowpack, releasing latent heat (USACE 1956; Colbeck 1975; Male and Gray 1981; DeWalle and

Rango 2008). The energy required to raise the temperature of a cold snowpack to 0°C is considerably lower than the energy required to initiate a phase transition from solid to liquid water (USACE 1956). Similarly, the heat energy supplied to the snowpack by latent heat of fusion is several times greater than sensible heat from rain. Therefore, rain falling on a sub-freezing snowpack could generate greater runoff compared with an equivalent rainfall event on a 0°C-isothermal snowpack. Snowmelt and rainfall may not result in runoff if the water freezes within the snowpack, while rainfall or meltwater on a warm snowpack can quickly translate into runoff (DeWalle and Rango 2008).

A number of factors influence the behaviour of rain or meltwater within a snowpack in addition to its temperature. The antecedent liquid water content of the snowpack, which can be up to 5% of the weight of the snowpack (Bruce and Clark 1966), affects the generation of runoff (Colbeck 1975; Garvelmann *et al.* 2015). A wet snowpack produces a faster runoff compared to a dry snowpack under equivalent rainfall (Colbeck 1975; Singh *et al.* 1997), as the dry snowpack retains rain or meltwater (Bruce and Clark 1966). Coastal mountain ranges in North America typically have a heavy, wet snowpack, while mountain ranges of interior climates, such as the Rocky Mountains, tend to have a thinner, dry snowpack (Mock and Birkeland 2000). Snow grain size, a function of the age of the snowpack, and therefore the degree of metamorphism that has occurred, also affects runoff lag times, where rain or meltwater will percolate through a large-grained snowpack more readily (Colbeck 1975). Additionally, water percolating through a snowpack affects the density (Colbeck 1975) and snow metamorphism within the snowpack (Singh *et al.* 1997; Hägeli and McClung 2003).

Numerous studies have examined runoff generation and flooding resulting from rain-on-snow events. For example, Garvelmann *et al.* (2015) found that 2-60% of the runoff generated during rain-on-snow events in the Black Forest region of Germany was attributed to snowmelt. Similarly, Sui and Koehler (2001) reported higher discharge during rainfall in winter compared with summer, due to the addition of snowmelt to runoff in southern Germany. Prowse and Owens (1982) found that snowmelt on the Craigieburn Range in New Zealand was greatest during days with rainfall due to high sensible heat transfer. In the Pacific Northwest, Mazurkiewicz *et al.* (2008) found that on average, rain-on-snow contributed to a small amount of meltwater annually; however, large events produced a high volume of runoff. Several severe winter flooding events attributed to heavy rainfall combined with snowmelt have occurred in California (Kattelmann *et al.* 1991; Kattelmann 1997), the Pacific Northwest (Marks *et al.* 1998; McCabe *et al.* 2007), and northeastern United States (Anderson and Larson 1996).

Given the importance of winter snowpack to water security and ecosystem services more broadly in western Canada, this research aims to increase the understanding of critical snowpack-sensitive climate variables. Additionally, evaluating trends in extreme hydrologic events is challenging due to the high-impact, low frequency nature of extreme events; however, evaluating the trends and variability of the drivers of hydrologic extremes provides fundamental insight into the potential generation of such extremes. Therefore, this research is focused on evaluating trends and variability in winter above-freezing air temperatures and associated rainfall. These hydroclimatic variables represent a critical threshold for the deterioration of winter snowpack, the potential generation of hydrologic extremes, such as river ice break-up and flooding, and are important indicators of changing hydroclimatic regimes in major watersheds in western Canada.

2.2. Study Area

Western Canada is characterized by mountains, valleys, and plateaus of the Western Cordillera, and the vast Prairie and Boreal forest regions to the east of the Rocky Mountains. The continental divide traverses the high-elevation peaks of the Rocky Mountains, separating west-, east-, and north-flowing rivers, and these alpine environments are important source regions for annual river flow (Figure 2-1). The Liard, Peace, and Athabasca rivers are major tributaries to the Arctic-flowing Mackenzie River. The Saskatchewan River flows east, contributing to the Nelson River and draining into Hudson Bay. The Fraser and Columbia river basins, covering the majority of southern BC, flow into the Pacific Ocean, along with relatively smaller river basins along the north coast of BC (the Stikine, Nass, and Skeena rivers). Peak streamflow on these rivers occurs during spring, driven by high-intensity snowmelt. Streamflow during the summer is generated through rainfall, glacier melt, and high-elevation snowmelt. Flow decreases during autumn and low flow is sustained throughout the winter.

These major rivers in western Canada span varied physiographic and hydroclimatic regions. The hypsometric profiles given in Figure 2-2 indicate the proportion of the watersheds located in alpine or low-elevation environments and describe the general topography of each basin (Strahler 1952), an important factor in snowmelt runoff (Wayand *et al.* 2015). The upper Peace, Athabasca, and Saskatchewan river basins are characterized by steep alpine headwaters and Rocky Mountain foothills with elevations above 1000 m, and a long, gradual slope through the middle and lower basins. The Peace and Liard rivers transition to a northern Boreal forest while the Athabasca and Saskatchewan rivers primarily flow over prairie grassland. Much of the upper Fraser, Columbia, and north coastal rivers are flanked by steep mountainous topography while only small portions of the lower basins of these rivers are below 1000 m.

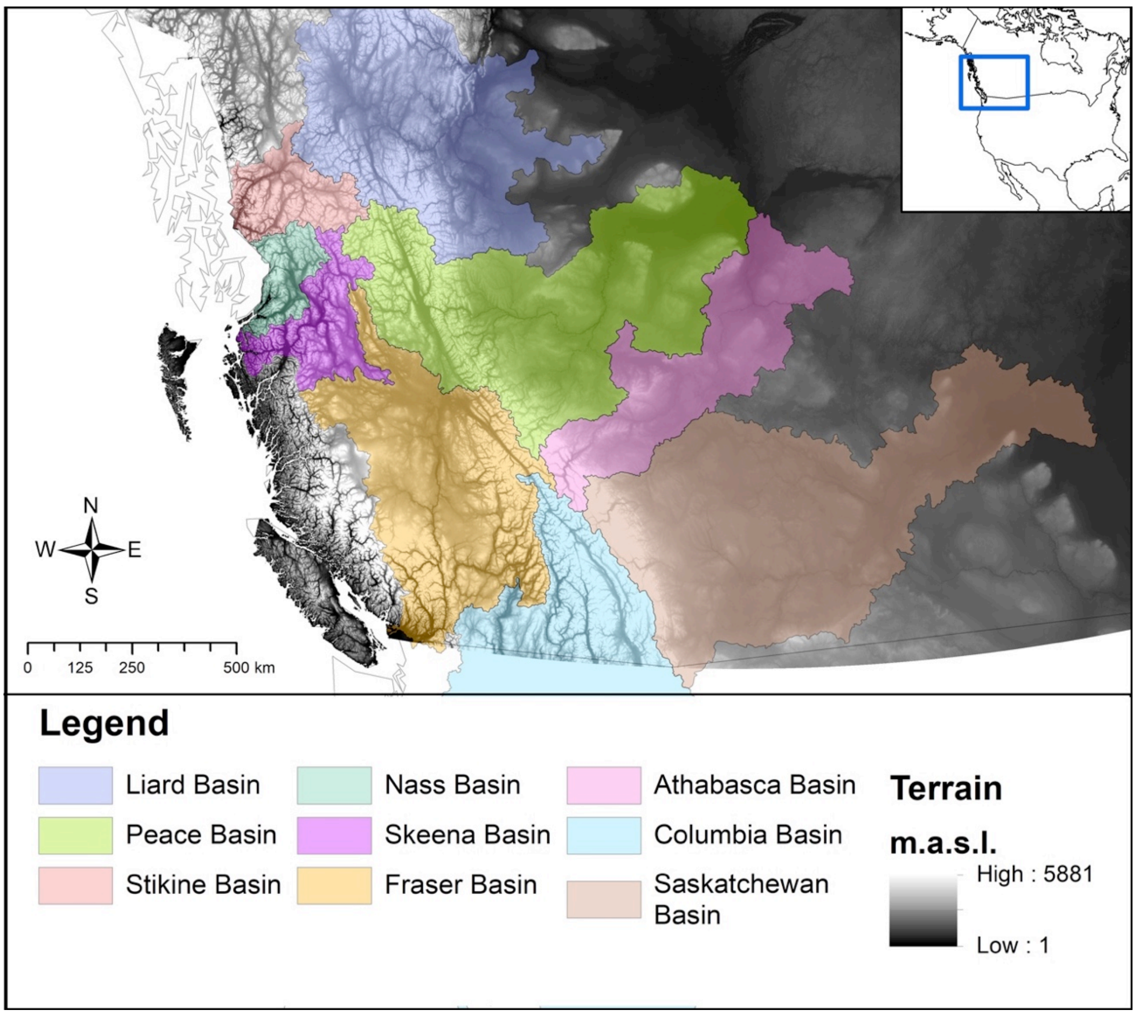

Figure 2-1: Major river basins in western Canada that are the focus of this research

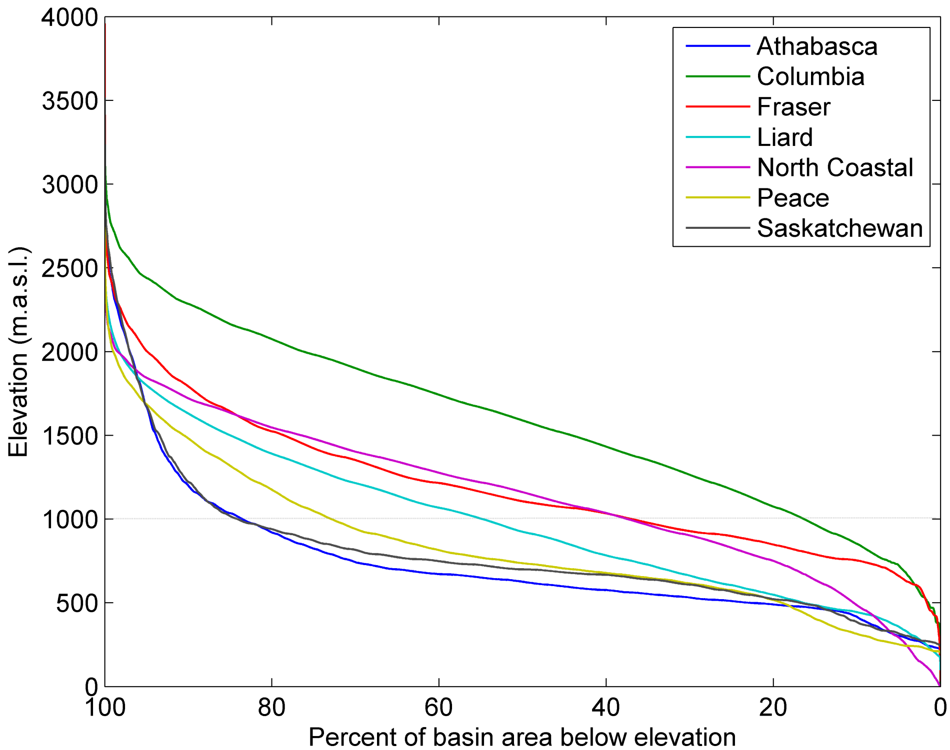

Figure 2-2: Hypsometric profiles of major river basins in western Canada

2.3. Data and Methods

High-resolution (1/16-degree) daily minimum and maximum air temperature (°C) and precipitation (mm) data for western Canada, from 1946 to 2012, are used to evaluate winter (DJFM) hydroclimatology (Werner *et al.* 2018). Although the existence and duration of snow-cover and/or below-freezing temperatures in western Canada is highly variable depending on latitude, elevation, and proximity to the coast, this definition of the winter season largely negates the effect of solar radiation on snowpack decay and provides temporal consistency for statistical analysis. A thin-plate spline interpolation of climate station data, including the adjusted and homogenized Canadian climate data (AHCCD; Mekis and Hogg 1999), global historic climatology network (GHCN; Vose *et* *al.* 1992), and United States historical climatology network (USHCN; Quinlan *et al.* 1987), using ClimateWNA climatology (Wang *et al.* 2012) as a covariate, was developed by the Pacific Climate Impacts Consortium (Werner *et al.* 2018). Daily mean temperature is calculated as the average value of the minimum and maximum temperature for that day (e.g. Klein Tank *et al.* 2002).

To evaluate surface air temperatures associated with snowmelt, the number of days when the mean daily temperature is above freezing (T_{mean} > 0°C) is calculated for every winter month (DJFM) and season from 1946-2012. Accumulated melting degree-days (MDD) are calculated as the sum of the mean daily temperatures above 0°C during each winter month and season for each year. Precipitation phase is determined by a defined threshold, or critical temperature, of 1°C. Days when mean daily temperature is equal to or above 1°C, precipitation is classified as rain, and below 1°C as snow. Precipitation phase is a function of numerous atmospheric and physical processes (Harpold *et al.* 2017) and at air temperatures nearing 0°C may fall as mixed rain and snow, slush, graupel, or hail, and the greatest uncertainty for precipitation phase classification exists when air temperatures are between 0°C and 2°C (Feiccabrino *et al.* 2012). A temperature threshold of 1°C is consistent with precipitation phase temperatures in similar regions (USACE 1956; Rohrer 1989; L'hôte *et al.* 2005; Yuter *et al.* 2006; Dai 2008; Lundquist *et al.* 2008; Kienzle 2008) and was found to greatly increase the accuracy of precipitation phase classification when compared with a 0°C threshold (Feiccabrino *et al.* 2013). A minimum daily precipitation of 2 mm is selected to remove minor rainfall amounts and account for data errors. Furthermore, runoff events are more likely to be triggered by rainfall events greater than 2 mm. Snowfall fraction is calculated as the percentage of total precipitation that falls as snow, as defined by the 1°C threshold.

Trends of surface climate variables are evaluated using the Mann-Kendall non-parametric test for trend (MK; Mann 1945; Kendall 1975) using a significance of $p < 0.05$. The magnitude of the trend is given by a linear slope calculated using Sen's method for slope estimation (Sen 1968). The MK test has been widely used to detect trends in hydrological and meteorological time series data (e.g., Burn 2008; O'Neil *et al.* 2017). Each grid point was evaluated for serial autocorrelation and as few grid points (< 10%) exhibited autocorrelation, "pre-whitening" (e.g., Yue *et al.* 2002) was not performed. Four points exhibiting a significant trend for all four variables, the frequency and magnitude of above-freezing temperatures and rainfall, are randomly selected from the headwaters of the Fraser, Peace, Athabasca, and Saskatchewan river basins to demonstrate seasonal variability.

2.4. Results

Winter climatology

The average temperature from December to March was calculated for every grid point in western Canada to identify areas that are below freezing during the winter season. This region, shown in blue in Figure 2-3, has a seasonal snow and river ice cover (Brown and Braaten 1998; Bennett and Prowse 2010; von de Wall 2011; Allchin and Déry 2017), and the colder regions, in darker shades of blue, are expected to have a sub-freezing snowpack throughout the winter. The coastal region of BC is predominantly above freezing and low-elevation river valleys in the Columbia, Fraser, and north coastal

(Stikine, Nass, and Skeena) river basins are temperate, remaining near or slightly below 0°C during winter, and, consequently, the snowpack in these basins is particularly vulnerable to temperature variability above and below freezing (Adam *et al.* 2009; Brown and Mote 2009). There is a noticeable contrast between the low- and high-elevation regions in the lower Fraser and Columbia river basins, evidence of the strong topographic gradients. The winter temperatures throughout much of the Liard, Peace, Athabasca, and Saskatchewan river basins, particularly the mountain headwaters and downstream reaches, are well below freezing throughout the winter season, suggesting a lengthy snow and ice cover season. The mean winter temperature in the north coastal, Fraser, and Columbia river basins decreases with increasing elevation (Figure 2-4). Conversely, the mean winter temperature in the Liard, Peace, Athabasca, and Saskatchewan river basins is coldest at low elevations and increases with increasing elevation to mid-elevations (1000-1500 m) then steadily decreases with increasing elevation.

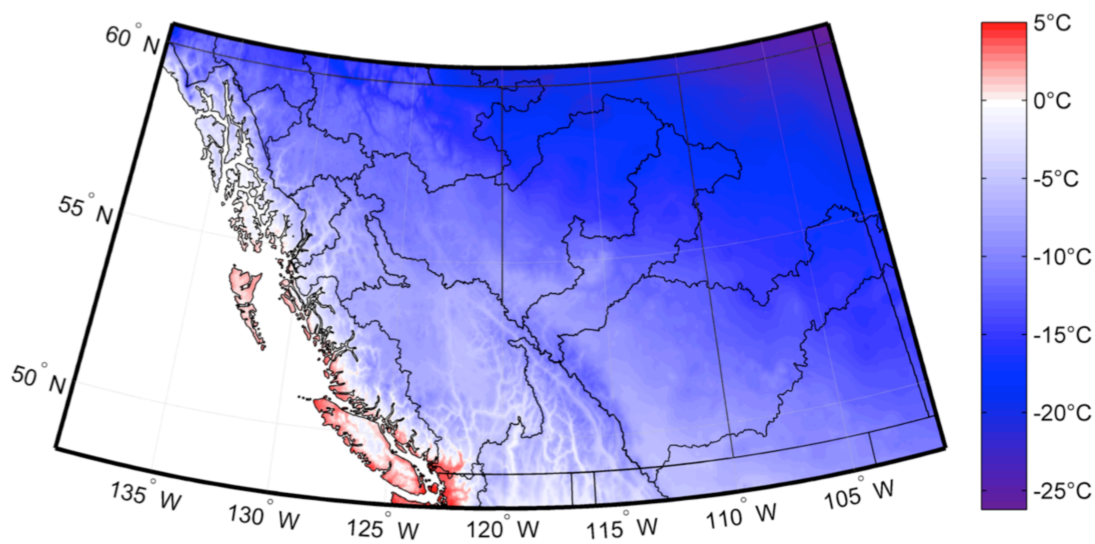

Figure 2-3: Average (1946-2012) winter temperature (DJFM) in western Canada

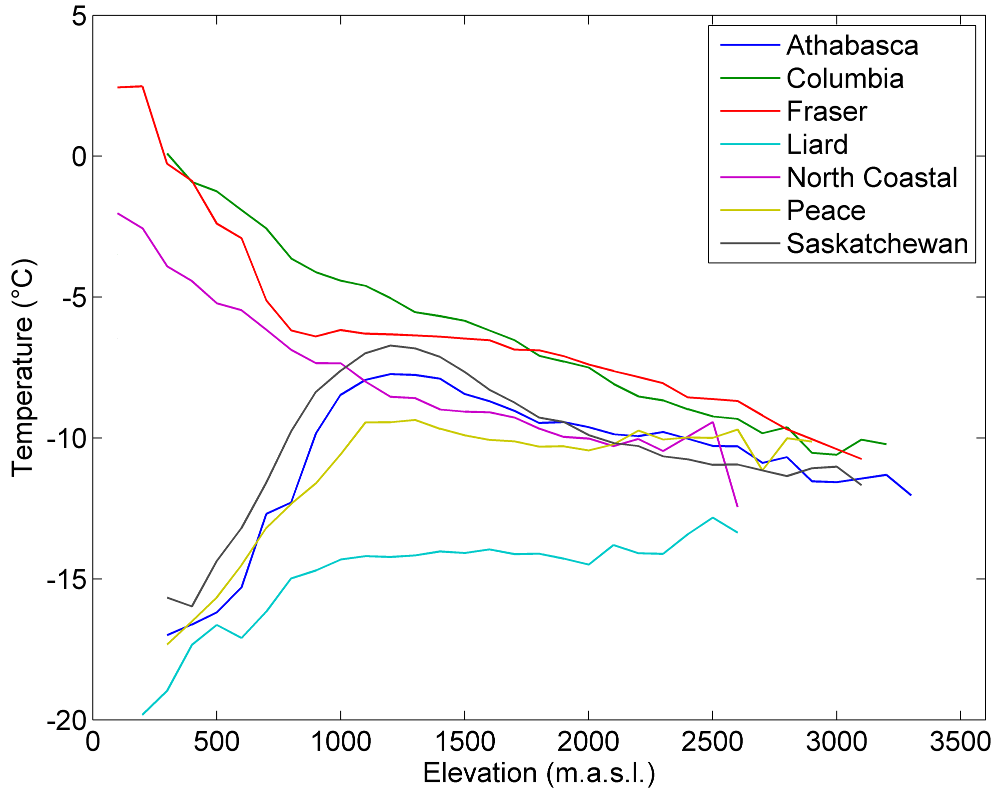

Figure 2-4 The distribution of mean winter (DJFM) temperature from 1946-2012, by 100 m elevation bands, in major river basins in western Canada

Mean cumulative winter (DJFM) precipitation (mm) was calculated for western Canada. Precipitation patterns are strongly influenced by both the topography of the study region and proximity to the Pacific Ocean. Precipitation is highest along the coast as incoming moist air masses impinge on the Coast Mountain range (Figure 2-5). For the Fraser, Columbia, and north coastal basins, precipitation is higher at low (below 500 m) and high (above 2000 m) elevations (Figure 2-6). Regions in a mountain rain shadow, including the Fraser Plateau in the central Fraser River basin and the Prairie and Boreal forest regions to the east of the Rocky Mountains receive relatively little precipitation. Precipitation in the Athabasca, Liard, Peace, and Saskatchewan river basins increases with increasing elevation (Figure 2-6). This relatively high precipitation on the Rocky and Columbia mountain ranges in southern BC and Alberta highlights the importance of the mountain snowpack in the headwaters of several major river basins in western Canada.

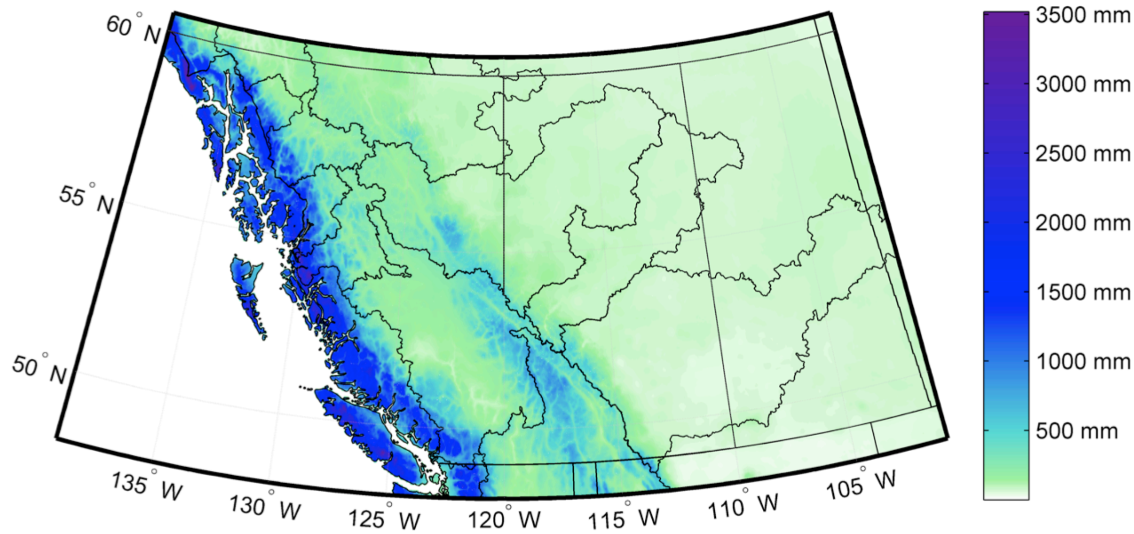

Figure 2-5: Mean cumulative winter (DJFM) precipitation (1946-2012) in western Canada

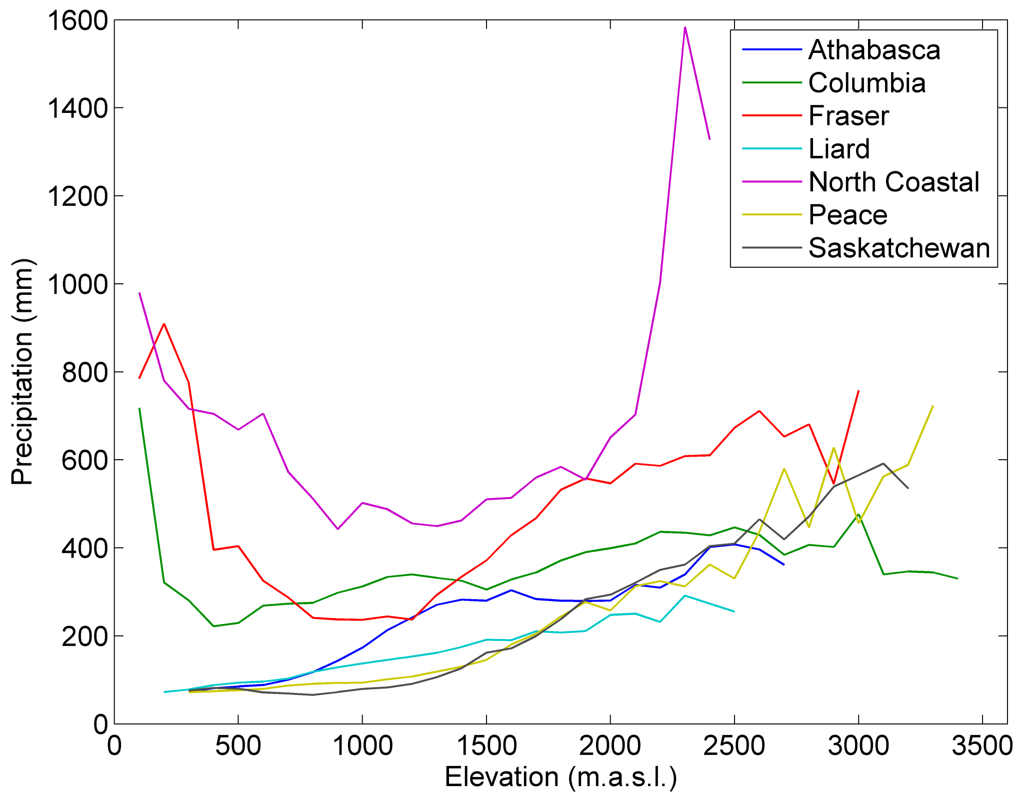

Figure 2-6: The distribution of mean winter (DJFM) precipitation from 1946-2012, by 100 m elevation bands, in major river basins in western Canada

Air Temperature Indicators

The mean number of winter (DJFM) days when the mean daily temperature is above freezing ($T_{mean} > 0°C$) is highest along the coast and in low-elevation river valleys in the Columbia and Fraser river basins (Figure 2-7). In the headwaters of the Columbia River basin there is a sharp contrast in the frequency of above-freezing temperatures between the Rocky Mountain Trench and surrounding mountains. The highest peaks of the Rocky Mountains along the southern BC-Alberta border remain below freezing throughout the winter, while the adjacent foothills region is prone to Chinook winds, contributing to the higher frequency of above-freezing temperatures (Longley 1967;

Goulding 1978). In portions of the upper Fraser, Peace and Athabasca river basins an average of 10-30% of winter days are above freezing, while the northern coastal basins, the Liard, and lower Peace, Athabasca, and Saskatchewan basins have a low average frequency (< 10%) of days when the mean daily temperature is above freezing.

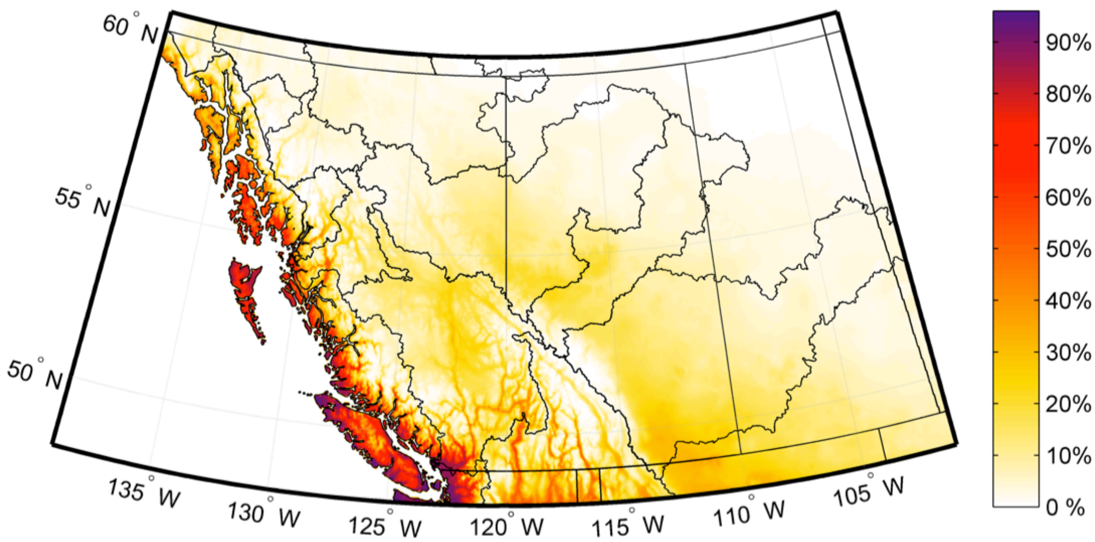

Figure 2-7: Average percentage of winter (DJFM) days when the mean daily temperature is above freezing from 1946-2012

It is apparent that the frequency of above-freezing temperatures during March exerts a strong influence on the total winter frequency of above-freezing days (Figure 2-8d). With the exception of the coastal region, most of the study region experiences very few above-freezing days in December (Figure 2-8a), January (Figure 2-8b), and February (Figure 2-8c), while the high frequency (> 60%) during March in the low-elevation areas of the Columbia and Fraser River basins indicates an early- to mid-March onset of snowmelt. In the upper Fraser, Peace, Athabasca, and Saskatchewan river basins 30-50% of days in March are above-freezing indicating a mid- to late-March onset of snowpack ripening or melt. High-elevation areas, including the mountainous headwater regions of the Saskatchewan, Athabasca, and Columbia rivers are predominantly below freezing throughout the winter.

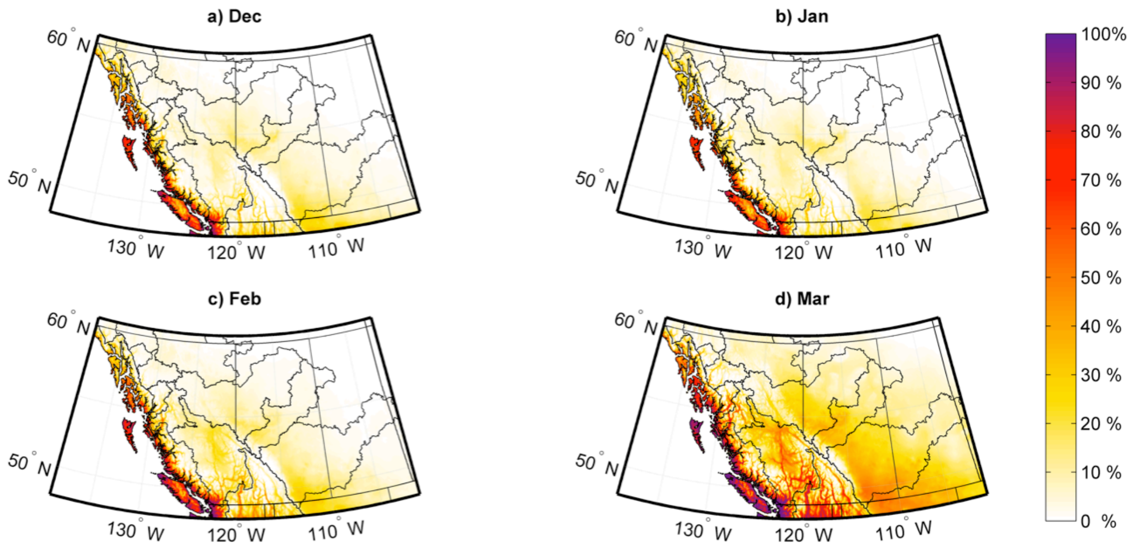

Figure 2-8: Average frequency, in percentage of days per month, when the mean daily temperature is above freezing for a) December, b) January, c) February, and d) March, from 1946-2012

Widespread increases are observed in the percentage of winter days per decade when the mean daily temperature is above freezing, and only those trends significant at $p < 0.05$ are shown (Figure 2-9). The highest magnitude trends (an increase of 3-4% per decade) are found in coastal regions, the lower Fraser River basin and low-elevation areas of the Columbia River basin. Moderate increases (1-2% per decade) are widespread across the Fraser, Peace, Athabasca, and Saskatchewan river basins. The time series at four points in the study region, in the headwaters of the Peace (1690 m), Fraser (845 m), Athabasca (1899 m), and Saskatchewan (1289 m) basins demonstrate considerable interannual variability superimposed on an overall increasing trend in the frequency of days when the mean daily temperature is above freezing (Figure 2-10). The greatest changes in the Fraser and north coastal basins have occurred below 500 m elevation, while the Athabasca, Saskatchewan, Peace, and Columbia basins are dominated by temperature increases between 1000 m and 1500 m elevation (Figure 2-11).

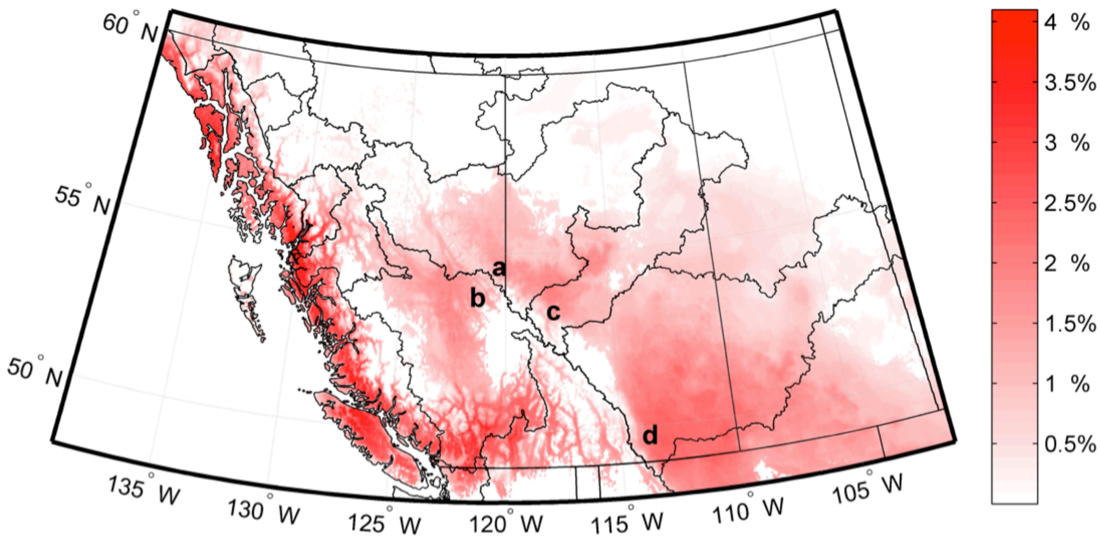

Figure 2-9: Trends in the percentage of winter (DJFM) days, per decade, when the mean daily temperature is above freezing from 1946-2012. Only those trends significant at p < 0.05 are shown

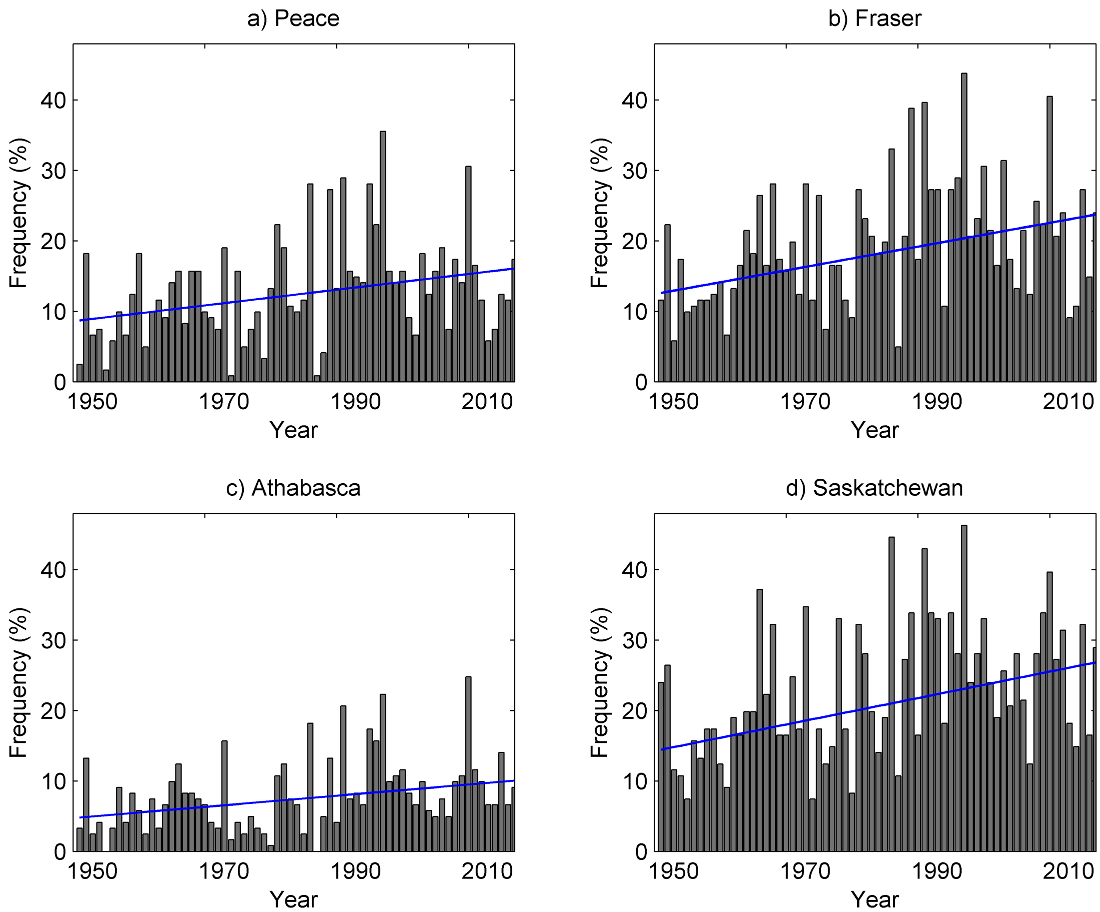

Figure 2-10: The time series of the percentage of winter days from 1946-2012 when the mean daily temperature is above freezing at four points in western Canada identified in Figure 2-9 – the headwaters of the a) Peace (1690 m), b) Fraser (845 m), c) Athabasca (1899 m), and d) Saskatchewan (1289 m) river basins. Trends (blue lines) are significant at $p < 0.05$

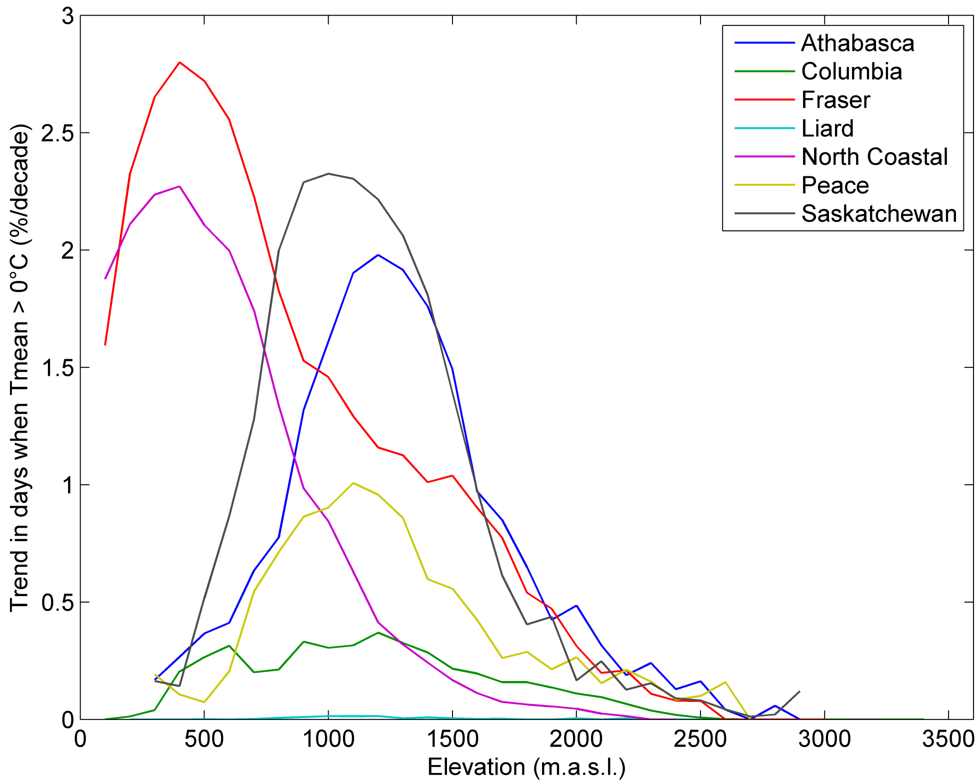

Figure 2-11: The distribution, by 100 m elevation bands, of significant ($p < 0.05$) trends in winter above-freezing temperatures in major river basins in western Canada from 1946-2012

The increases in winter above-freezing days are dominated by strong, significant ($p < 0.05$) trends during March, particularly in the Saskatchewan, Athabasca, Fraser, and Columbia river basins (Figure 2-12d). These trends are indicative of an earlier onset of snowpack ripening or snowmelt, and are consistent with trends toward an earlier spring freshet reported by Whitfield (2001), Burn (2008), Rood *et al.* (2008), and Kang *et al.* (2016). Minimal increases along the coast, and decreases in the Columbia River basin are seen during December (Figure 2-12a), and minimal, patchy trends exist during February in the Fraser, Athabasca, and Saskatchewan river basins (Figure 2-12c). Widespread increases of up to 2% per decade occurred during January in the Fraser River basin and the upper Peace, Athabasca, and Saskatchewan river basins (Figure 2-12b).

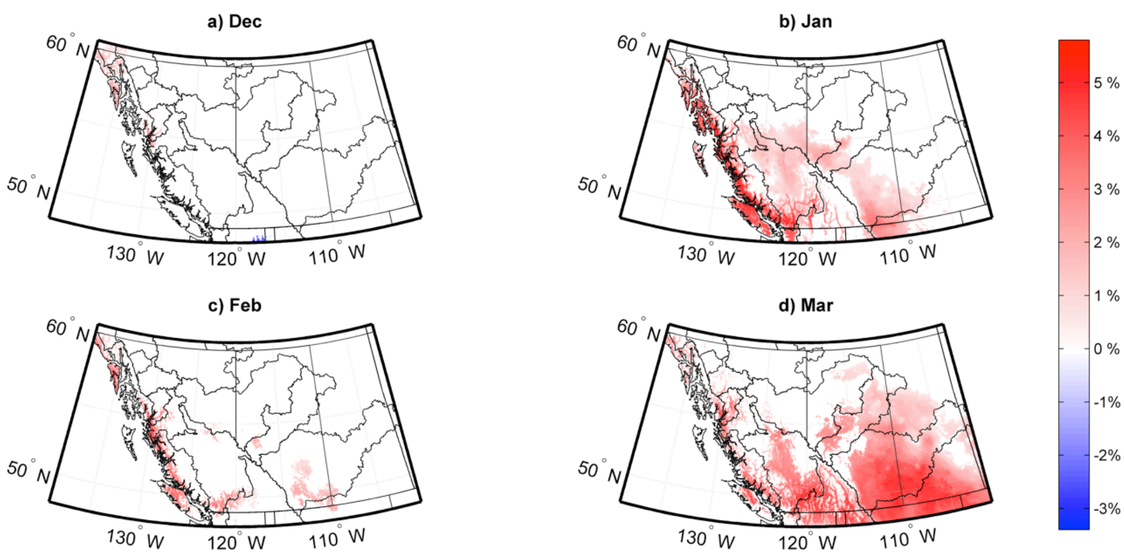

Figure 2-12: Trends in the percentage of days per decade when the mean daily temperature is above freezing for a) December, b) January, c) February, and d) March, from 1946-2012. Only trends significant at $p < 0.05$ are shown

Winter accumulated melting degree-days (MDD) are indicative of the heat energy available to the snowpack during the winter. Spatial patterns of accumulated MDD are consistent with the frequency of above-freezing days (Figure 2-13); however, the magnitude of accumulated MDD suggests that many winter days are above, but still close to freezing. Accumulated MDD are highest along the coast and low-elevation river valleys in the Fraser and Columbia river basins, and lowest in high-elevation areas including the Rocky, Columbia, and Coast Mountains, in the Liard, north coastal, and the lower Peace, Athabasca, and Saskatchewan river basins. Patterns of accumulated MDD throughout the winter are dominated by higher temperatures during March (Figure 2-

14d). Small regions of higher accumulated MDD are evident in the upper Fraser, Peace, Athabasca, and Saskatchewan river basins during December, January, and February (Figure 2-14a-c), indicating a higher potential for snowmelt and runoff generation. The relatively high accumulated MDD adjacent to and east of the Rocky Mountains, extending from the Peace to the Saskatchewan basins may be related to the prevalence of Chinook winds in the region (Longley 1967; Goulding 1978).

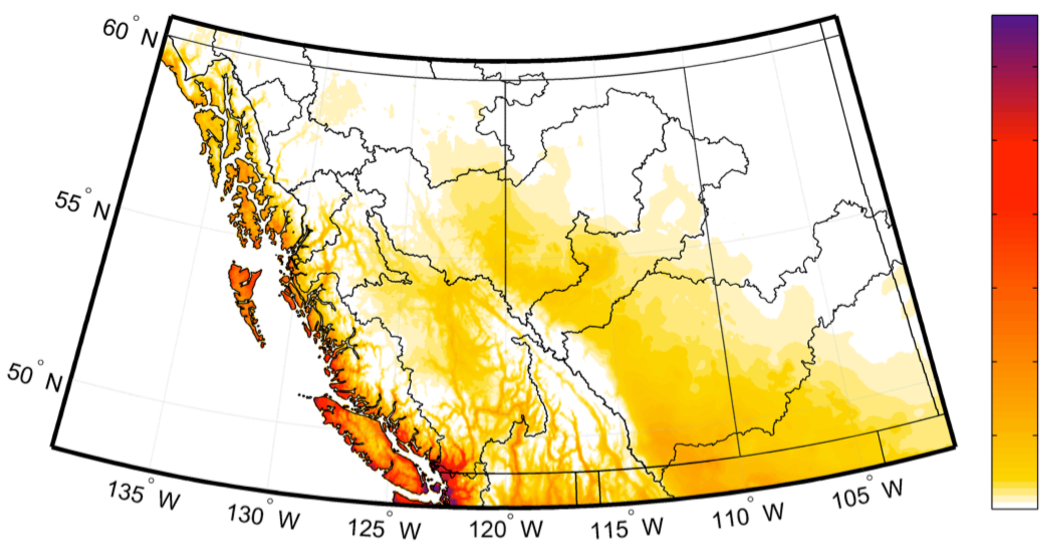

Figure 2-13: Average winter (DJFM) accumulated melting degree-days (MDD) from 1946-2012

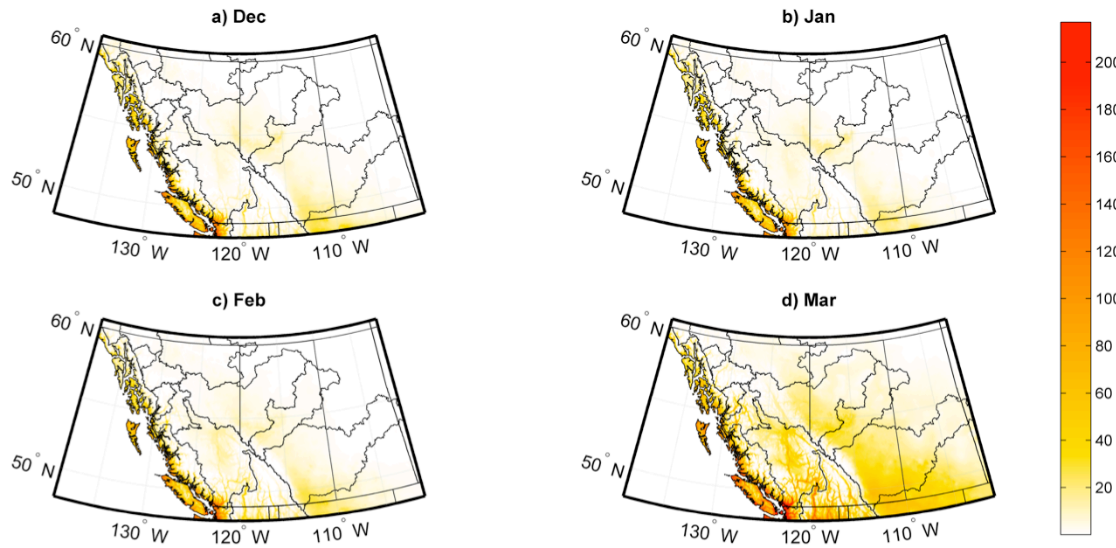

Figure 2-14: Average accumulated melting degree days (MDD) per month for a) December, b) January, c) February, and d) March, from 1946-2012

Significant ($p < 0.05$) increases in winter (DJFM) accumulated MDD are prevalent along the coast and in the Saskatchewan River basin, portions of the upper Athabasca and Peace river basins, and in lower-elevation areas of the Fraser River basin (Figure 2-15). The time series of the four points in the study region - in the headwater regions of the Peace, Fraser, Athabasca, and Saskatchewan river basins, show substantial interannual variability in winter accumulated MDD, particularly in the latter half of the study period, and noticeable spikes in 1992 and 2005 in the Peace, Athabasca, and Fraser river basins, and in 1981, 1986, 1992, and 2005 in the Saskatchewan River basin (Figure 2-16). Increases in accumulated MDD in the Fraser and north coastal basins are strongest at elevations below 500 m while in the Peace, Athabasca, and Saskatchewan basins the greatest changes are seen between 1000 and 1500 m elevation (Figure 2-17).

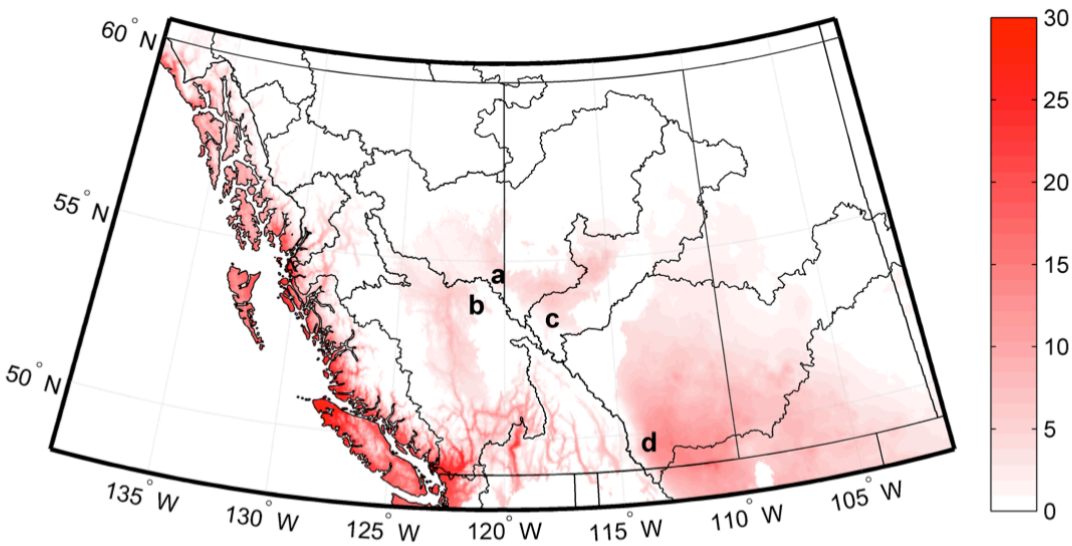

Figure 2-15: Trends in accumulated melting degree-days, given per decade from 1946-2012. Only trends significant at $p < 0.05$ are shown

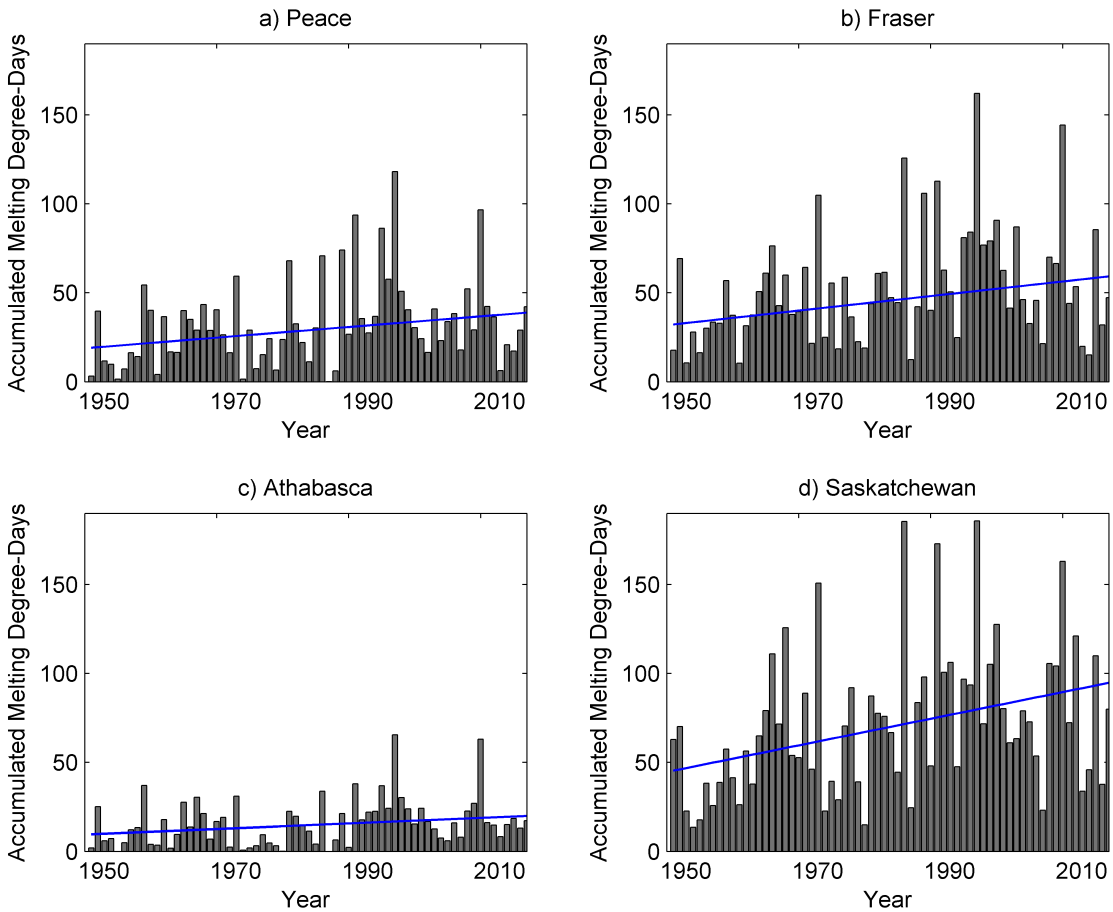

Figure 2-16: The time series of accumulated melting degree days from 1946-2012 at four points in western Canada identified in Figure 2-15 – the headwaters of the a) Peace (1690 m), b) Fraser (845 m), c) Athabasca (1899 m), and d) Saskatchewan (1289 m) river basins. Trends (blue lines) are significant at $p < 0.05$

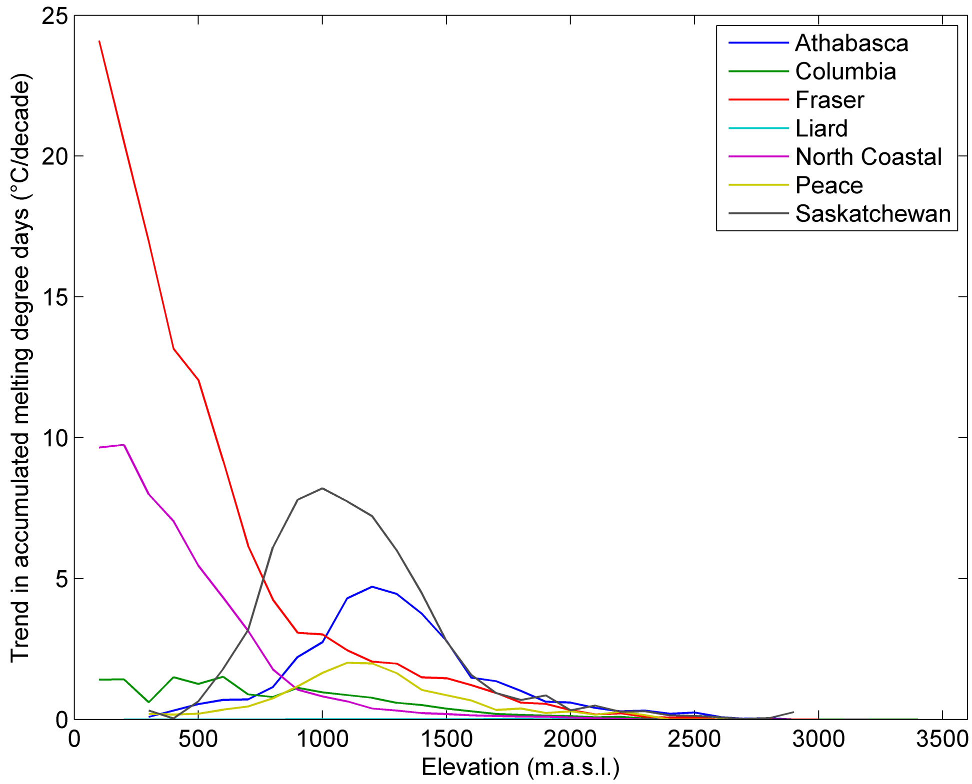

Figure 2-17: The distribution, by 100 m elevation bands, of significant ($p < 0.05$) trends in accumulated melting degree-days in major river basins in western Canada from 1946-2012

The magnitude of the winter accumulated MDD trends in the Fraser and Saskatchewan river basins are strongly influenced by increasing trends during March (Figure 2-18d). Significant ($p < 0.05$), but low-magnitude trends are evident in the upper Saskatchewan, Athabasca, Peace, and Fraser river basins during January (Figure 2-18b), while trends are largely nonexistent during December and February (Figure 2-18a,c).

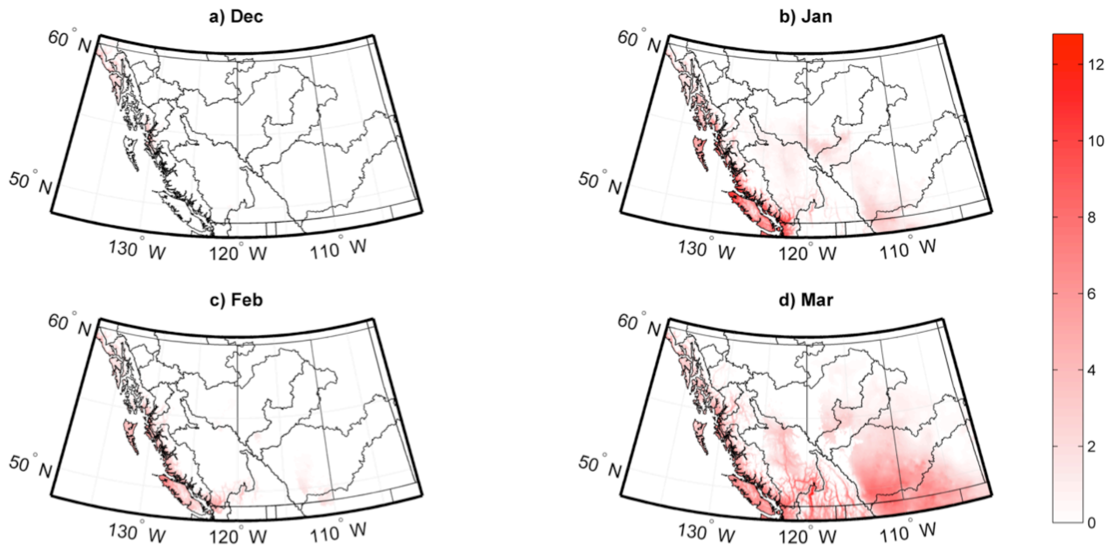

Figure 2-18: Trends in accumulated melting degree days per decade for a) December, b) January, c) February, and d) March, from 1946-2012. Only trends significant at $p < 0.05$ are shown

Precipitation Indicators

The number of days when rainfall occurs was evaluated using a temperature-index equation with a temperature threshold of $T_{mean} \geq 1°C$ and a minimum precipitation of 2 mm. The frequency is highest along the coast where rainfall occurs during more than 50% of winter days (Figure 2-19). Low-elevation river valleys of the Columbia and Fraser basins have a higher frequency of rainfall, while the upper Peace, upper Athabasca, and upper Saskatchewan see a very low frequency (< 5%) of days with rainfall.

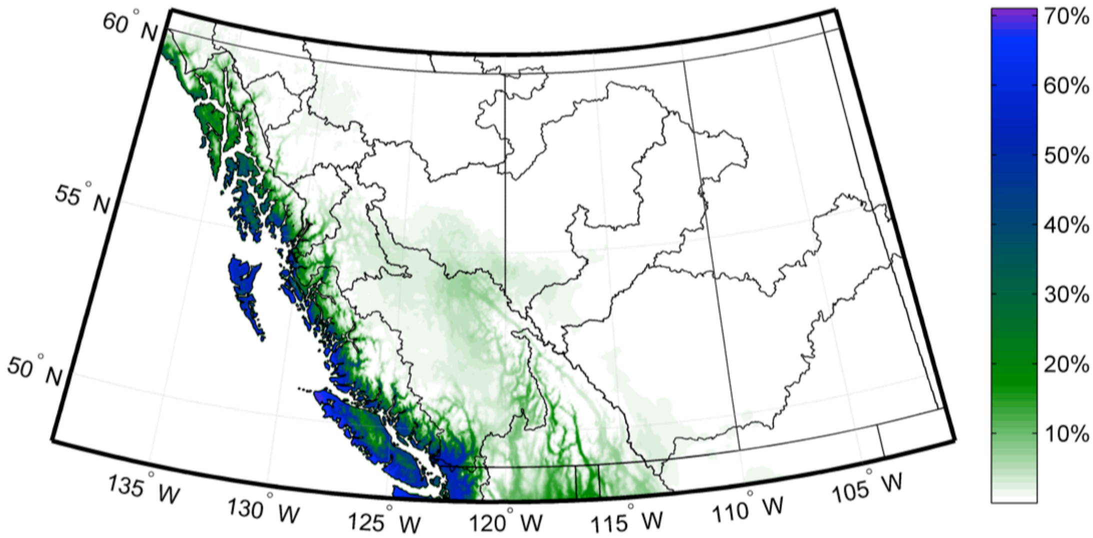

Figure 2-19: Average frequency, in percentage of days per winter, when a minimum of 2 mm of rainfall occurs from 1946-2012

Winter rainfall occurs along the coast during every month of the winter season (Figure 2-20a-d), while the Columbia, Fraser, Nass, Skeena, upper Peace and upper Athabasca primarily see rainfall during March (Figure 2-20d). The Fraser, Skeena, and upper Peace river basins receive rain during all winter months with an elevated frequency during March (Figure 2-20a-d). The frequency of winter days with rainfall remains low in the area to the east of the Rocky Mountains, consistent with low overall precipitation for this region. The percentage of winter days that are above freezing and also accompanied by rainfall is given in Figure 2-21, revealing a clear topographic delineation between high (> 40%), moderate (20-40%) and low (< 20%) co-occurrence of above-freezing temperatures and rainfall in the study region.

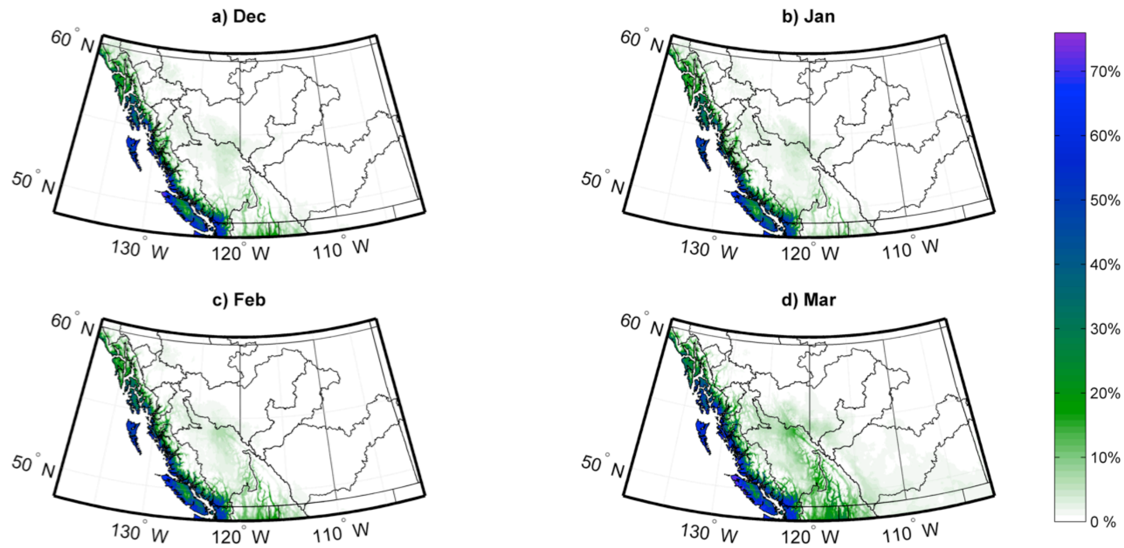

Figure 2-20: Average percentage of days when rainfall occurs for a) December, b) January, c) February, and d) March, from 1946-2012

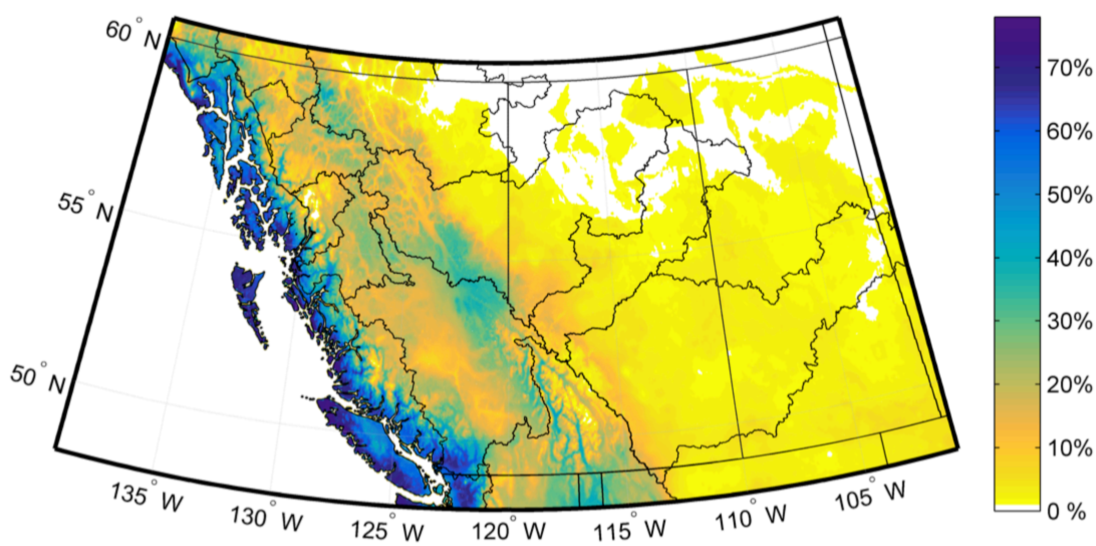

Figure 2-21: Percentage of winter days that are above freezing and also accompanied by rainfall from 1946-2012

There are statistically significant ($p < 0.05$) increases in winter days when rainfall occurs along the coast, the lower and eastern Fraser River basin, low-elevation valleys of the Columbia basin, and a portion of the upper Peace and Athabasca river basins (Figure 2-22). The time series at the four points in the study area are indicative of high interannual variability in the upper Fraser and upper Peace basins and moderate variability in the upper Athabasca and upper Saskatchewan basins (Figure 2-23). In the Fraser River basin, the greatest increases are seen below 500 m, with moderate increases between 1000 and 2000 m elevation (Figure 2-24). Similarly, increases in the north coastal basins peak below 500 m, while the greatest trends in the Columbia River basin occur between 500 and 1500 m, in the Peace River basin between 1000 and 2400 m, and in the Athabasca and Saskatchewan river basins between 1200 and 2000 m elevation.

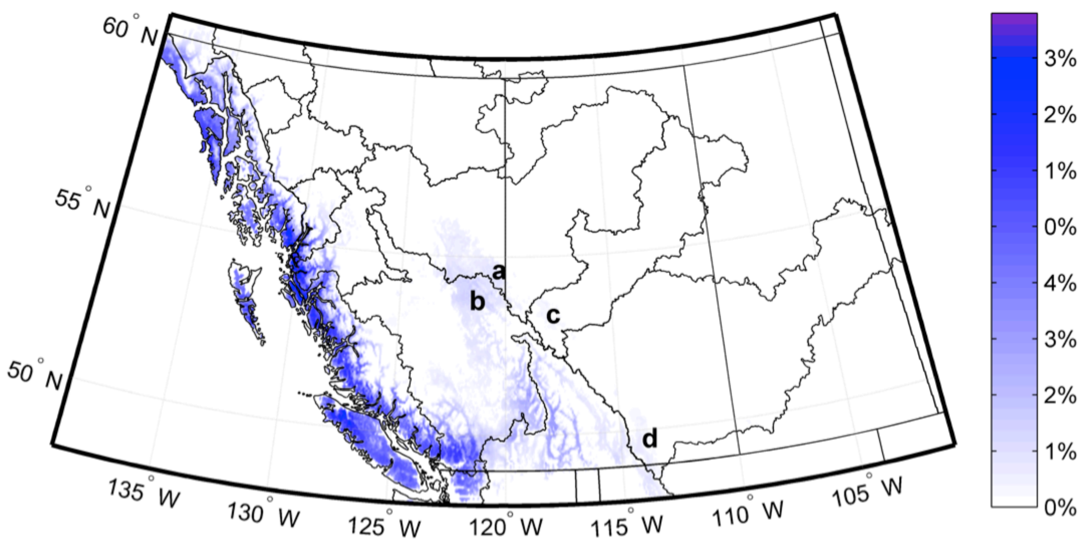

Figure 2-22: Trends in the percentage of winter days when rainfall has occurred, given per decade, from 1946-2012. Only trends significant at $p < 0.05$ are shown

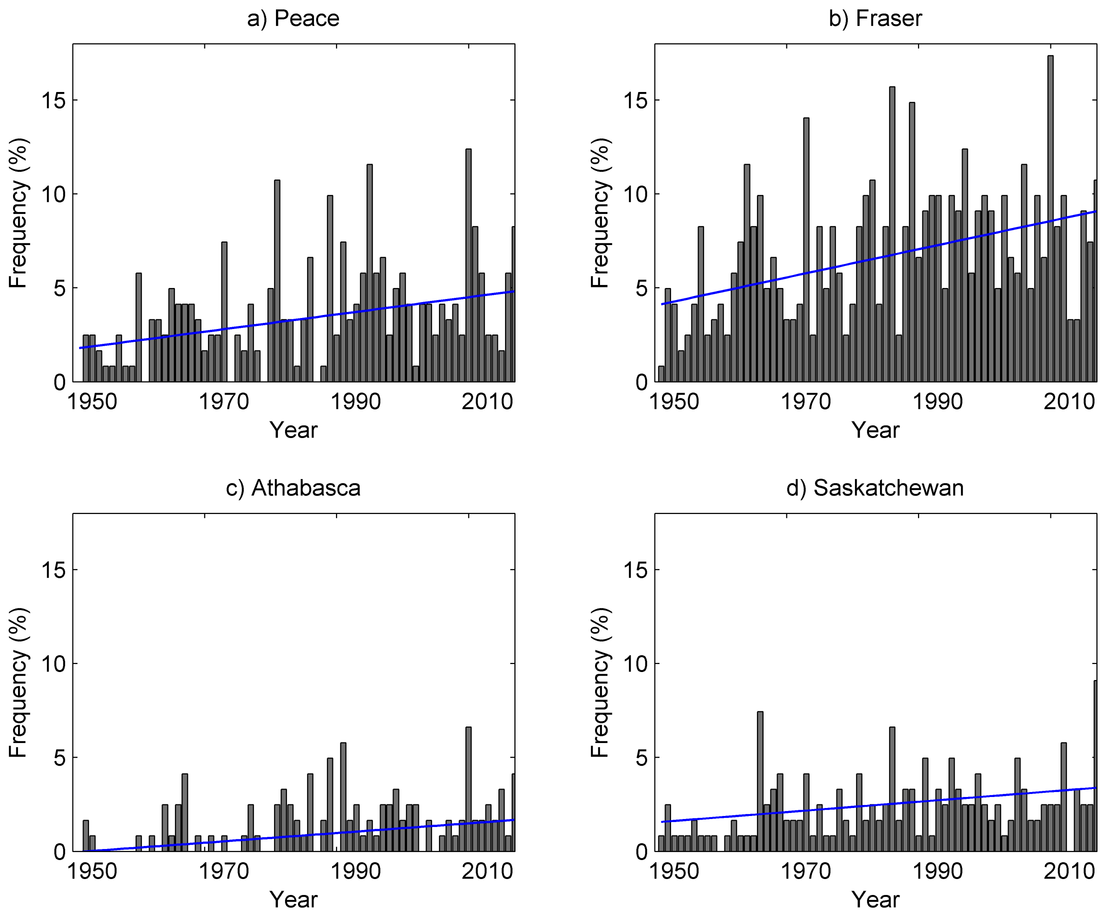

Figure 2-23: The time series of the frequency of winter days when rainfall has occurred from 1946-2012 at four points in western Canada identified in Figure 2-22 – the headwaters of the a) Peace (1690 m), b) Fraser (845 m), c) Athabasca (1899 m), and d) Saskatchewan (1289 m) river basins. Trends (blue lines) are significant at $p < 0.05$

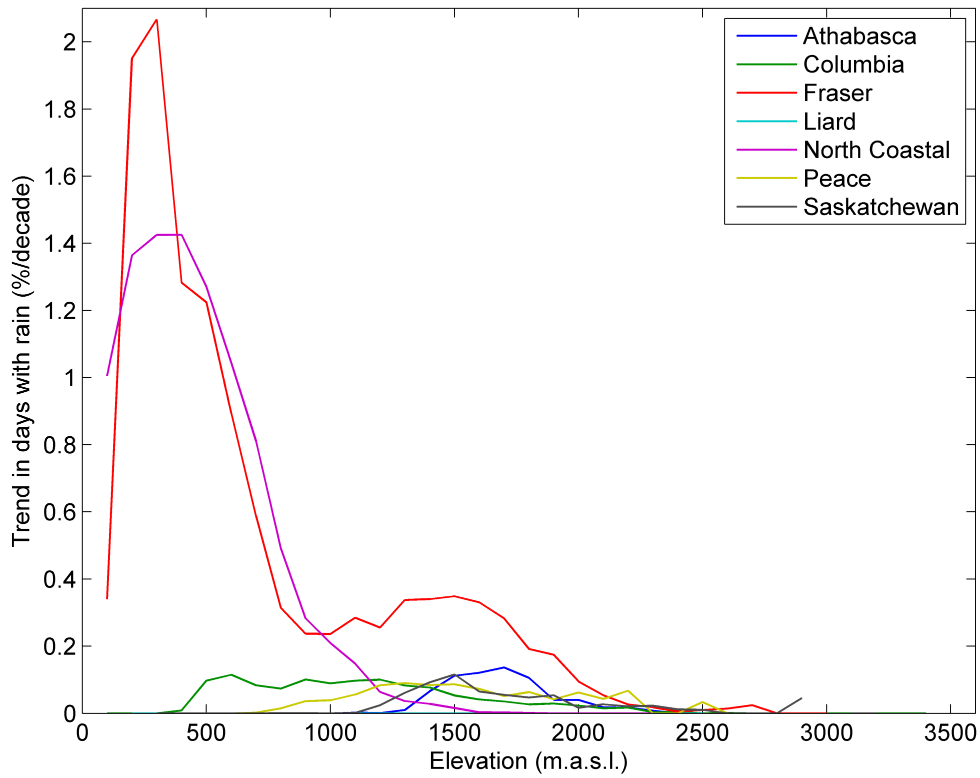

Figure 2-24: The distribution, by 100 m elevation bands, of significant ($p < 0.05$) trends in the frequency of winter days when rainfall has occurred in major river basins in western Canada from 1946-2012

The increase in days when winter rainfall occurs along the coast primarily occurs during January and March (Figure 2-25b,d), with minimal increases during December and February (Figure 2-25a,c). Increases in the interior basins primarily occur during March (Figure 2-25d) and are consistent with decreasing spring SWE, snow cover extent, and earlier onset of snowmelt (Brown and Mote 2009; Hernández-Henríquez *et al.* 2015; Kang *et al.* 2016; Najafi *et al.* 2017). A small area of the Fraser River near the outlet has a decreasing trend in winter days with rainfall in December and February; however, large increases in January and March result in an overall increase during the winter.

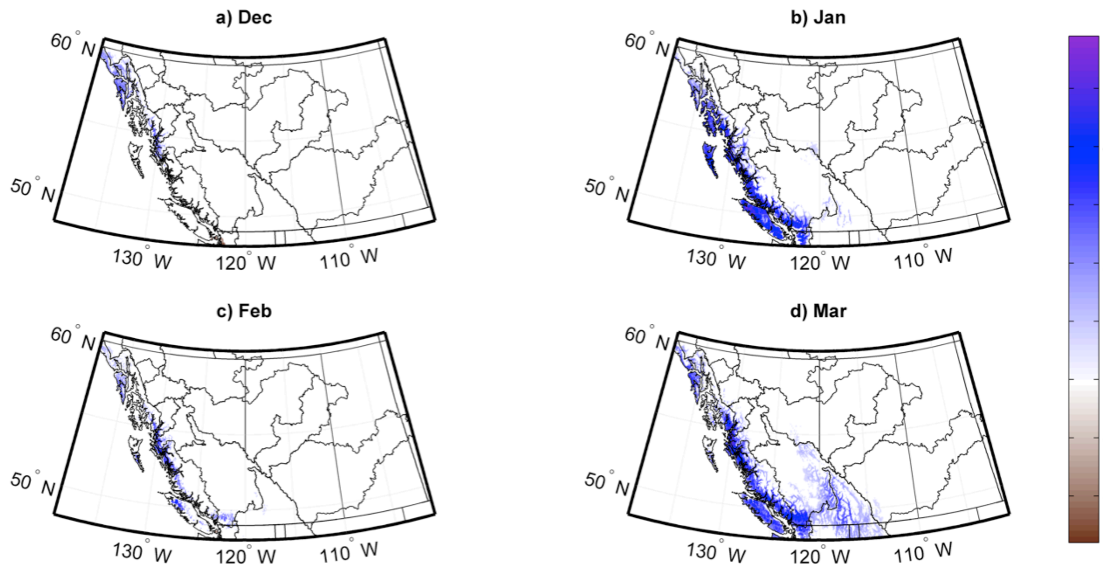

Figure 2-25: Trends in percentage of days per decade when rainfall has occurred for a) December, b) January, c) February, and d) March, from 1946-2012. Only trends significant at $p < 0.05$ are shown

The mean winter rainfall is very high along the coast while the Nass, Skeena, Fraser, and upper Peace, receive an average of 0-50 mm of rainfall during the winter (Figure 2-26). The majority of the rainfall in the interior basins occurs during March (Figure 2-27d). Smaller amounts of rain in a small region of the upper Fraser and upper Peace basins and in the low-elevation valleys of the Columbia basin occur December through February (Figure 2-27a-c).

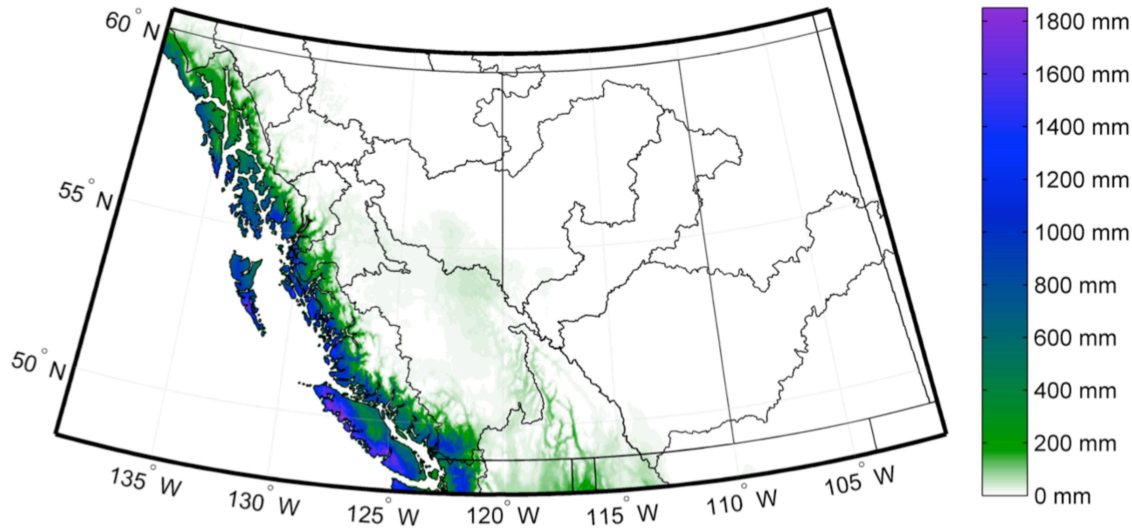

Figure 2-26: Average winter (DJFM) rainfall (mm) from 1946-2012

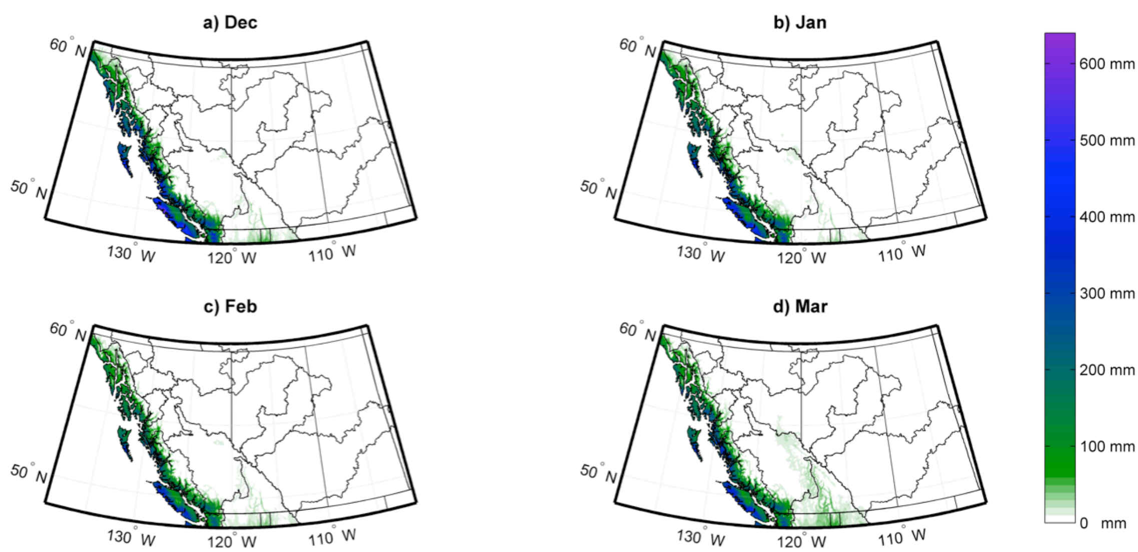

Figure 2-27: Average rainfall (mm) during a) December, b) January, c) February, and d) March, from 1946-2012

Patterns of statistically significant ($p < 0.05$) increases in winter rainfall are similar, but less pronounced than the spatial patterns of days when rainfall occurs (Figure

2-28). Increases have occurred along the coast, the lower and eastern areas of the Fraser and Columbia river basins, and a portion of the upper Peace and upper Athabasca river basins. The time series plots of the four points in the headwaters of the Peace, Fraser, Athabasca, and Saskatchewan river basins demonstrate the large interannual variability of rainfall superimposed upon overall significantly ($p < 0.05$) increasing trends, particularly in the upper Fraser and upper Peace basins (Figure 2-29). Increases in the north coastal basins primarily occur below 1000 m, while the distribution of increases in the Fraser basin is bimodal, with peaks centred at 500 m and 1500 m elevation, occurring at the lower and upper ends of the basin (Figure 2-30). The greatest rate of change in the Peace, Athabasca, and Saskatchewan basins occurs between 1200 and 2300 m, 1600 and 2200 m, and 1400 and 2700 m, respectively.

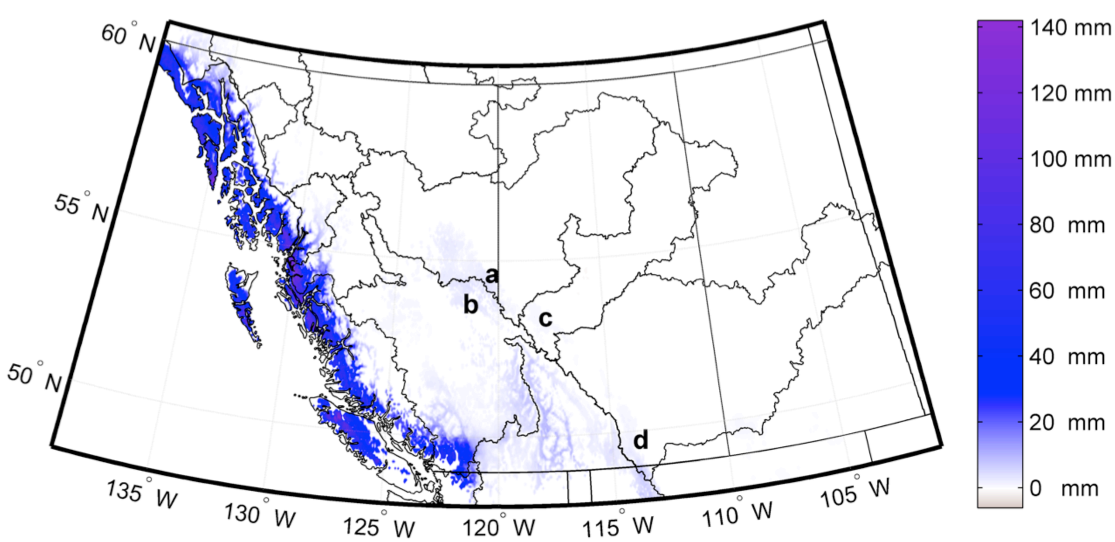

Figure 2-28: Trends in the amount of winter rainfall, given per decade, from 1946-2012. Only trends significant at $p < 0.05$ are shown

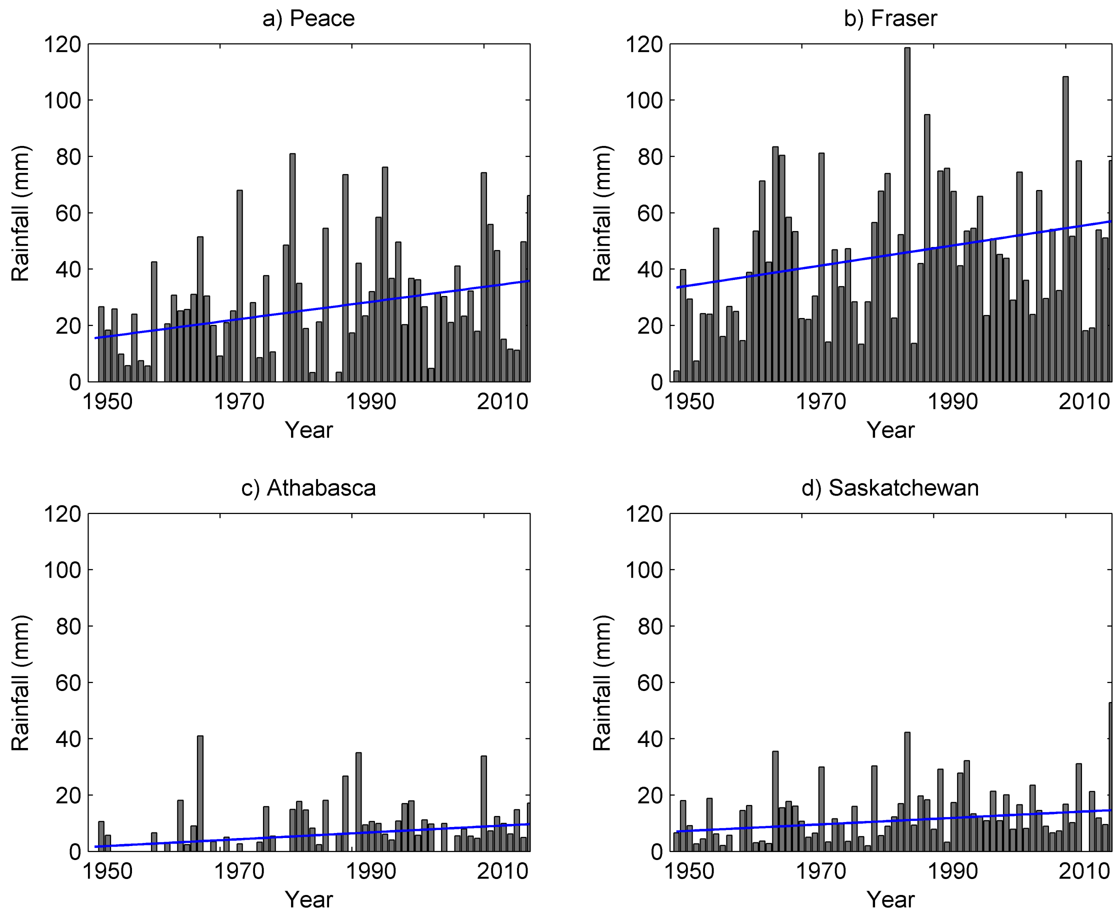

Figure 2-29: The time series of winter rainfall from 1946-2012 at four points in western Canada identified in Figure 2-28 – the headwaters of the Peace (1690 m), Fraser (845 m), Athabasca (1899 m), and Saskatchewan (1289 m) river basins. Trends (blue lines) are significant at $p < 0.05$

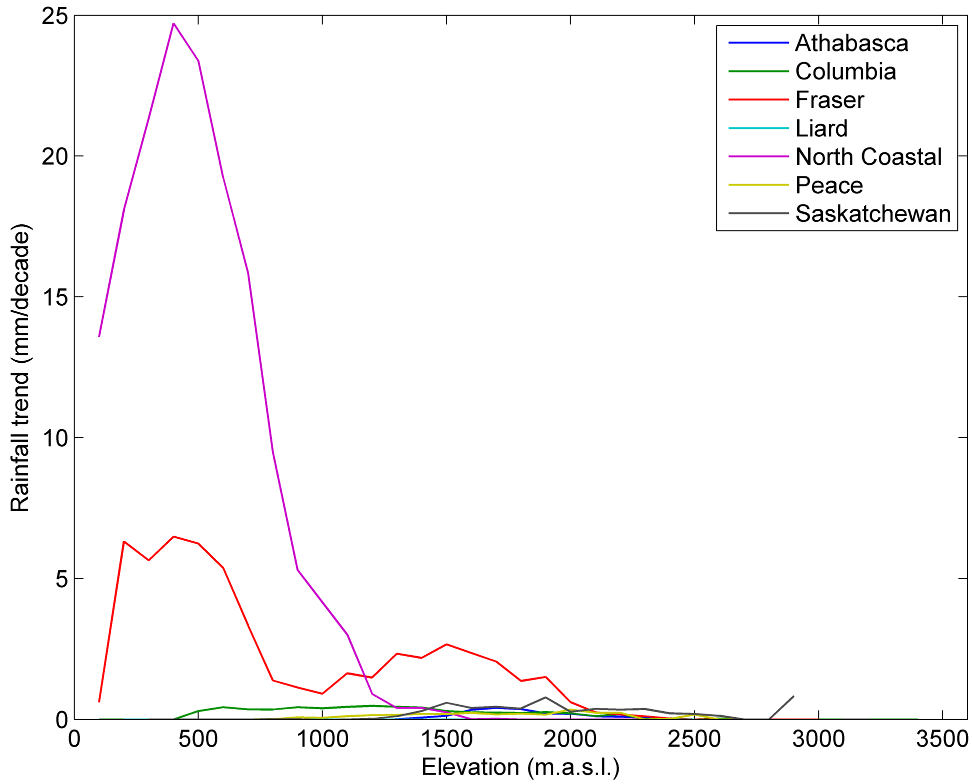

Figure 2-30: The distribution, by 100 m elevation bands, of significant ($p < 0.05$) trends in winter rainfall amounts in major river basins in western Canada from 1946-2012

The increases in winter rainfall along the coast are dominated by January and March rainfall trends (Figure 2-31b,d). Increasing trends in the Fraser, Columbia, and small portions of the headwaters of the Peace and Saskatchewan river basins are relatively minor and are largely confined to March (Figure 2-31d), although some increasing trends are evident in January in the headwaters of the Peace and Fraser river basins (Figure 2-31b). Increases in the frequency and amount of rainfall in these basins increases the risk of mid-winter or early spring flood generation and river ice break-up.

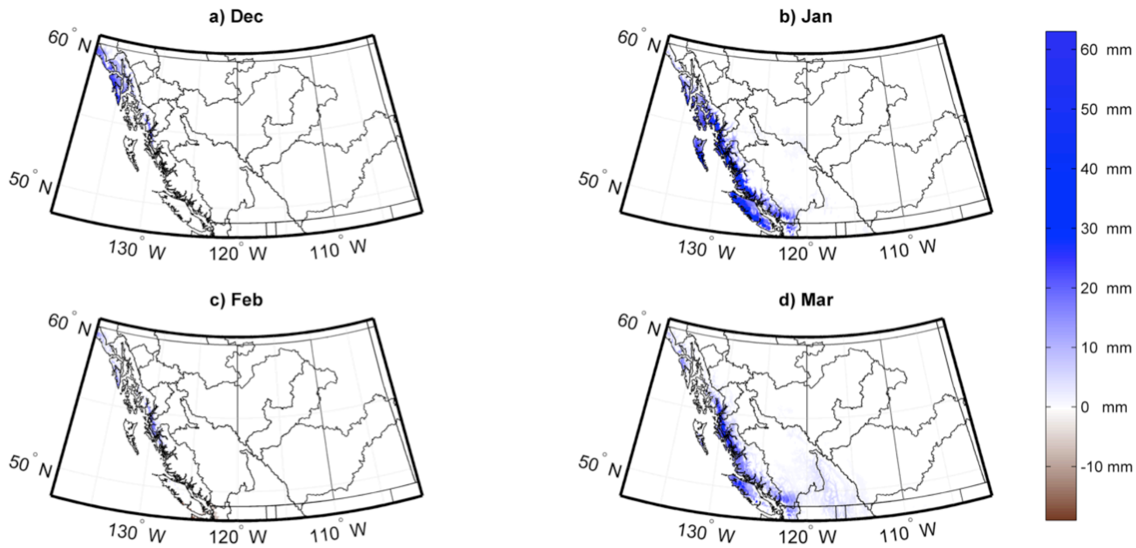

Figure 2-31: Trends in the amount of rainfall in a) December, b) January, c) February, and d) March, from 1946-2012. Only trends significant at p < 0.05 are shown

Winter (DJFM) snowfall fraction, or the percentage of precipitation falling as snow, is very high throughout most of the study area, with the exception of the coastal region (Figure 2-32). Despite a wide band of very high precipitation along the coast, only the nearshore regions have a very low snowfall fraction, indicating high snow accumulation in the Coast Mountains. Approximately 50% snowfall and rainfall is seen in the low-elevation regions of the Columbia River basin in southern BC and northern Washington State. Snowfall fraction is high (80-90%) in the Liard, Peace, Fraser, Athabasca, and Saskatchewan river basins. Statistically significant ($p < 0.05$) trends in snowfall fraction are predominantly decreasing and spatially diverse (Figure 2-33). The greatest decrease in snowfall fraction occurs along the coast, with moderate decreases (1-2% per decade) in the headwaters of the Peace and Fraser river basins, and stronger trends (3-4% per decade) in the lower Fraser and upper Columbia river basins.

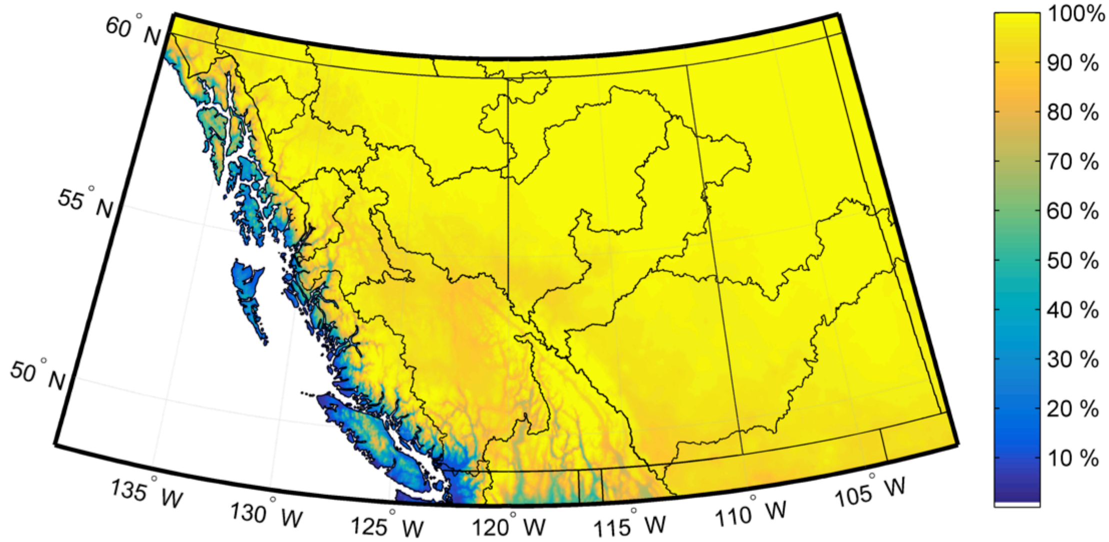

Figure 2-32: Average winter snowfall fraction, calculated as the percentage of total winter (DJFM) precipitation that falls as snow from 1946-2012

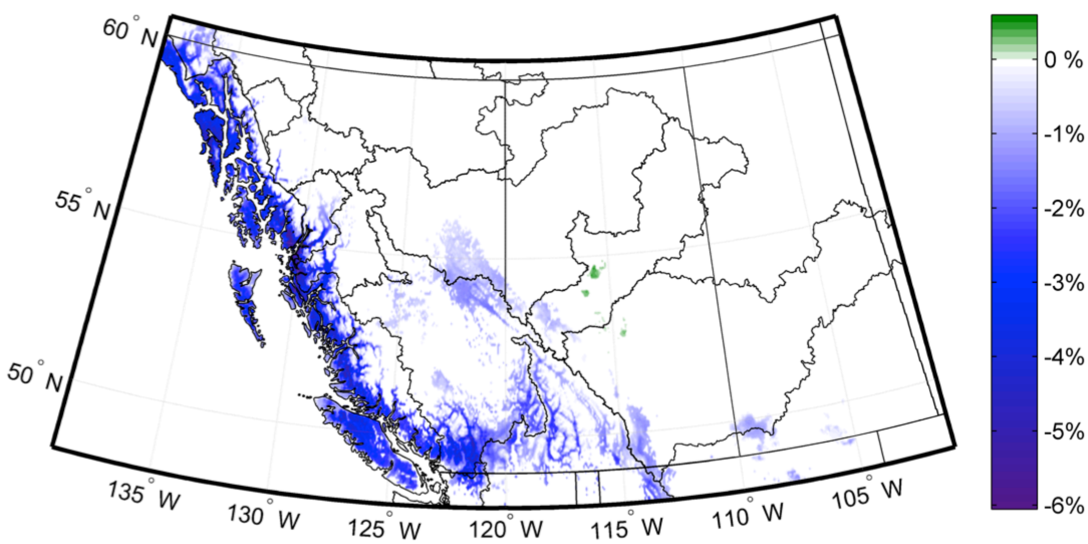

Figure 2-33: Trends in snowfall fraction (%) per decade from 1946-2012. Only trends significant at $p < 0.05$ are shown

2.5. Discussion and Conclusions

This study evaluated the prevalence and trends in winter temperature and precipitation over western Canada conducive to snowpack depletion, runoff generation, and extreme hydrologic impacts such as flooding and river ice break-up. Extreme hydrological events are hazardous and can have devastating impacts to biophysical and socio-economic systems. Furthermore, rivers in western Canada are highly dependent upon winter snow accumulation and spring melt for warm season water resource availability (Sauchyn and Kulshreshtha 2008; Walker and Sydneysmith 2008). Of the two primary concerns, the generation of an extreme hydrologic event is an immediate high-impact consequence and the deterioration of winter snowpack has seasonal and annual implications for water availability in western Canada. Results indicate an overall trend toward an increased frequency and magnitude of above-freezing temperatures and rainfall in western Canada; however, the trends vary spatially, and at different elevations across the study region, giving rise to critical elevation zones that pose the greatest threat to snowpack integrity and seasonal runoff.

Accumulated MDDs are indicative of heat energy available to raise the temperature of a cold (< 0°C) snowpack and/or melt a 0°C-isothermal snowpack. Similarly, the number of winter days when the mean daily temperature is above freezing signifies the frequency of days when heat energy is available to the snowpack. The greatest rate of increase in the frequency and magnitude of above-freezing temperatures in the north coastal and Fraser river basins is seen at low- to mid-elevations (below 1000 m), with increases occurring up to 2500 m elevation while increases in the Columbia basin are greatest between 500 and 1500 m. Additionally, winter rainfall increases are predominant in the Fraser River basin below 700 m and between 1200-1900 m, below 1000 m in the north coastal basins, and 500-1500 m in the Columbia River basin. The bimodal nature of the rainfall trends in the Fraser River basin follows the distribution of precipitation by elevation bands, suggesting trends in rainfall are a function of the amount of precipitation at various elevation bands up to 1900 m, above which the temperatures are below freezing and precipitation primarily falls as snow. The distribution of precipitation in the north coastal basins is also higher at low and high elevations; however, the lack of above-freezing temperatures at these elevations results in precipitation falling as snow.

Increases in above-freezing temperatures and rainfall in the Fraser and Columbia river basins are dominated by strong increasing trends during March, indicative of an earlier onset of spring snowmelt and freshet. This is consistent with previous research that has revealed decreasing spring SWE (Kang *et al.* 2014; Kang *et al.* 2016; Najafi *et al.* 2017), snow cover extent (Déry and Brown 2007; Brown and Mote 2009) and an earlier spring freshet (Whitfield 2001; Zhang *et al.* 2001; Stewart *et al.* 2005). Furthermore, snowfall fraction has significantly decreased in portions of the Fraser and Columbia basins, consistent with Vincent and Mekis (2006) and Vincent *et al.* (2015) who reported widespread declines in snowfall fraction in western Canada. Increases in above-freezing temperatures in the Columbia, Fraser, and north coastal basins are also seen in January, consistent with Linton (2014) who found significant increase in January snowmelt from 1950-2010. These trends raise concerns over implications for mid-winter snowpack integrity, SWE, and the generation of hydrologic extremes, suggesting that these events are more likely to occur during January compared with December and

February. This is reflected in a study by Newton *et al.* (2017) who identified 18 mid-winter river ice break-up events in the Fraser and Columbia river basins during January between 1950 and 2008, while fewer events occurred during December, February, and March.

The snowpack in the north coastal, Fraser, and Columbia river basins is influenced by the proximity to the Pacific Ocean and is typically a wet snowpack (Mock and Birkeland 2000) and, therefore, above-freezing temperatures and rainfall are expected to result in a rapid runoff (Colbeck 1975; Singh *et al.* 1997). Numerous winter flooding events have been triggered by rain-on-snow in British Columbia (Eaton *et al.* 2002) and the Pacific Northwest region of the United States (Marks *et al.* 1998; McCabe *et al.* 2007). Rain-on-snow resulted in weak layers in the snowpack that were conducive to avalanche activity in the Columbia Mountains (Fitzharris 1987; Hägeli and McClung 2003). Furthermore, these basins are located in a temperate region, and are particularly vulnerable to shifts in temperature fluctuations as the mean winter temperature is just below 0°C (Adam *et al.* 2009; Brown and Mote 2009; Kerkhoven and Gan 2011). Loss of snowpack at these elevations may threaten water availability for agricultural productivity (Kerkhoven and Gan 2011; Coelho 2015) and the generation of hydroelectricity in the Columbia River basin (Cohen *et al.* 2000) and on the Nechako River in the upper Fraser basin (Déry *et al.* 2012) and may pose challenges for water resource management (Payne *et al.* 2004). The trends identified in this research indicate an increased risk for the generation of extreme hydrologic events and reduced snowpack in the north coastal, Fraser, and Columbia river basins that may be exacerbated under continued climate change.

The greatest rates of increase in the frequency and magnitude of above-freezing temperatures in the Peace, Athabasca, and Saskatchewan river basins have occurred at low to mid-elevations (700-1500 m) in the foothills and Rocky Mountains headwater regions. These elevations have the highest average winter temperature in the Peace, Athabasca, and Saskatchewan river basins. The mean winter temperature decreases gradually above 1500 m, suggesting that under continued climate change the high-elevation areas of these river basins will be vulnerable to above-freezing winter temperatures. There are minor increases in rainfall in these basins; however, the increases are primarily seen at higher elevations (1500-2200 m). This is in agreement with Kerkhoven and Gan (2011) who reported that snowpack in the Athabasca River basin is more sensitive to changes in temperature compared to precipitation. The majority of annual streamflow on these rivers originates as snowpack in the mountains and foothills (Ashmore and Church 2001); therefore, increases in temperature and precipitation conducive to mid-winter or premature snowmelt at these elevations, and decreasing snowfall fraction in the headwaters of these basins, are particularly concerning for warm season water availability. Patterns of increasing above-freezing temperatures and rainfall are consistent with the patterns of declining snowfall fraction in the headwaters of the Peace, Athabasca, and Saskatchewan river basins.

The increases in above-freezing temperatures are dominated by increases during January in the mountain headwaters and foothills regions of the Peace, Athabasca, and Saskatchewan. During March increasing trends are widespread in the Athabasca and Saskatchewan basins indicating a trend towards an extensive loss of snowpack in the southern Alberta Prairies and early onset of snowmelt in the mountain headwaters. The loss of snow cover in the Prairies has implications for infiltration and groundwater recharge (Hayashi and van der Kamp 2005). Moderate increases in winter rainfall have occurred in the upper Peace basin, adjacent to an area of increased rainfall in the upper Fraser basin, particularly during January and March. Winter precipitation, including rainfall, is relatively low east of the Rocky Mountains, where above-freezing temperatures are the prominent factor leading to loss of snowpack and the generation of hydrologic extremes including numerous mid-winter river ice break-up events (Newton *et al.* 2017). Increases in above-freezing temperatures and rainfall in the Peace River basin are evident in the mountain headwaters raising concerns over the loss of snowpack. These results are consistent with Romolo *et al.* (2006b) and O'Neil *et al.* (2017) who found that snowmelt was occurring significantly earlier in the Peace basin and Najafi *et al.* (2017) who reported declining spring SWE in the upper Peace basin.

Mid-winter snowmelt has the potential to generate hydrologic extremes and enhanced streamflow on a tributary may trigger a river ice break-up event on a larger river (Prowse 1986; Peters and Prowse 2001). The loss of water storage in snowpack threatens water supply for the generation of hydroelectricity on the Peace River (Peters and Prowse 2001) and several tributaries to the Saskatchewan River (Martz *et al.* 2007), and reduces water availability for agricultural and municipal use (Martz *et al.* 2007; Pentney and Ohrn 2008) and downstream ecosystems (Prowse *et al.* 2006). For example, high streamflow resulting from high-intensity spring snowmelt and river ice jams are essential for downstream flooding, nutrient replenishment, and biodiversity in the Peace-Athabasca Delta (Peters *et al.* 2006). The role of reduced mountain snowpack on summer drought in the Prairies has been documented (e.g., Bonsal *et al.* 2011; Hanesiak *et al.*

2011); however, it is unknown to what extent mid-winter or premature spring snowmelt has on drought. There is considerable interannual variability in snowpack magnitude, and the combination of low winter snow accumulation and mid-winter melt events could result in anomalously low snowpack, spring freshet volume, and summer water availability, increasing the need for enhanced water resource management strategies.

Snowpack in the Rocky Mountain headwaters of the Peace, Athabasca, and Saskatchewan rivers is typically dry (Mock and Birkeland 2009), which tends to retain meltwater (Bruce and Clark 1966); therefore, runoff response is not expected to be rapid and greater energy inputs may be required to generate runoff compared with a dense, wet snowpack. A period of above-freezing temperatures or rainfall can accelerate snow metamorphism and alter the structure of the snowpack (Colbeck 1975; Singh *et al.* 1997; Hägeli and McClung 2003) or retain meltwater (Bruce and Clark 1966). Depending on the temperature of the snowpack, there may be a greater threshold for energy inputs to trigger snowmelt. Future research should compare periods of above-freezing temperatures and rainfall on snow survey measurements and streamflow in the mountain headwaters of these basins to determine the hydrologic impacts of these hydroclimatic conditions and potential thresholds under given snowpack conditions.

Trends in winter above-freezing temperatures and rainfall identified in this research suggest the need for further comprehensive analysis, such as identifying the large-scale atmospheric drivers of these hydroclimatic conditions. Results confirm that snowmelt-generating hydroclimatic conditions are a feature of the winter snow climatology throughout western Canada during winter. Furthermore, spatially variable patterns of increasing above-freezing temperatures, accumulated MDD, days with rainfall, and total rainfall indicate an increasing risk of hydroclimatic extremes. Results of this study provide a basis for understanding the generation of winter intra-seasonal hydroclimatic extremes occurring well in advance of the spring melt, particularly in critical zones of major watersheds. Future research should focus on basin-scale processes and quantify hydrologic response to these hydroclimatic conditions including comparison with snow survey and hydrometric data.

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Chapter 3: Atmospheric drivers of winter above-freezing temperatures and associated rainfall in western Canada

Abstract

Winter snowpack is an important water storage component critical for water resource availability in major watersheds in western Canada. Temperature and precipitation are key drivers of snowmelt, and are strongly influenced by large-scale atmospheric circulation. Daily winter synoptic-scale mid-tropospheric circulation patterns from 1949-2012 are classified into 12 dominant types using Self-Organizing Maps (SOM). Corresponding patterns of above-freezing temperatures and rainfall are identified using a daily, high-resolution gridded dataset to understand characteristics of atmospheric circulation conducive to loss of snowpack and the generation of extreme hydrologic events. A ridge of high pressure over western Canada is associated with a high frequency of above-freezing temperatures and the strength of the surface climate response is related to the strength and position of the ridge. Conversely, a ridge of high pressure over the Pacific Ocean and adjacent trough of low pressure over western Canada, and zonal flow are associated with a low frequency of above-freezing temperatures. Rainfall is most frequent and widespread on days when there is a mid-tropospheric ridge of high pressure over western Canada, and is primarily seen in along the coast and in the upper Peace, Fraser, and Columbia river basins. Two synoptic types were found to have a step-change in mean frequency and slope in 1977, one depicting zonal flow (Type 3; decrease) and one characterized by a strong ridge of high pressure centred near the coast (Type 10; increase), coinciding with a shift to a positive phase of the Pacific Decadal Oscillation. The 12 synoptic types are further grouped into four circulation regimes representing

zonal flow, a ridge of high pressure over the Pacific Ocean, a ridge of high pressure over western Canada, and a ridge of high pressure centred near the west coast, to facilitate analysis of the role of persistence in surface climate response. Winters dominated by a persistent ridge of high pressure over western Canada have moderate to high accumulated melting degree days and low to moderate rainfall compared to winters dominated by zonal flow or a ridge of high pressure over the Pacific Ocean and are expected to have a thinner snowpack and higher probability of an extreme hydrologic event. Results of this research have provided new insight into the role of the atmospheric circulation, particularly related to frequency and persistence, in driving winter hydroclimatic variability conducive to snowmelt.

3.1. Introduction

The availability of freshwater in western Canada is dependent upon winter snowpack, particularly in the mountain headwater regions of major river basins (Barnett *et al.* 2005; Stewart 2009). Spring freshet, typically the largest hydrologic event in cold regions, is dominated by snowmelt, while high-elevation snowpack contributes to streamflow throughout the summer. A deficit of water resources during critical periods threatens numerous biophysical and socio-economic systems, including agricultural productivity (Pentney and Ohrn 2008), hydroelectricity generation (Filion 2000; Roberts *et al.* 2006), and aquatic ecosystems (Wrona *et al.* 2006; Burn *et al.* 2008; Wrona *et al.* 2016). Furthermore, a lack of adequate snow accumulation or anomalously early melt can contribute to summer drought conditions (Bonsal *et al.* 2011; Hanesiak *et al.* 2011). Many of the recent changes to climate have affected the magnitude of winter snow accumulation and timing of melt, raising concerns over diminishing water security (Walker and Sydneysmith 2008; Sauchyn and Kulshreshtha 2008) and the potential for extreme hydrologic events (Bates *et al.* 2008), such as river ice break-up (Beltaos 2002; Newton *et al.* 2017) and flooding (Anderson and Larson 1996; Marks *et al.* 1998; McCabe *et al.* 2007). Therefore, there is a growing need to understand the complex drivers of climatic and hydrologic variability to effectively inform water resource management (McGregor 2017).

The magnitude of snow accumulation is a function of hydroclimate throughout the winter season. The onset of snow accumulation and melt are strongly associated with air temperatures falling below and rising above freezing (Bonsal and Prowse 2003; Brown and Mote 2009), while end-of-season snow water equivalent (SWE) is related to the amount and phase of cold season precipitation and any mid-winter melt events. There is considerable hydroclimatic variability across western Canada, both spatially and temporally (Whitfield and Cannon 2000; Zhang *et al.* 2000; Edwards *et al.* 2008; Shabbar *et al.* 2011; Vincent *et al.* 2015; Edwards *et al.* 2017). Snow accumulation has decreased and snowmelt has occurred earlier in western Canada over recent decades. For example, O'Neil *et al.* (2017a) quantified reductions in snow accumulation and timing of melt in major river basins in western Canada using high-resolution gridded climate data, from 1950-2010, and a temperature-index snow accumulation and melt model and found widespread declines in both snow accumulation and melt. Kang *et al.* (2016) found that snowpack in the Fraser River basin had declined between 1949 and 2006, and the snowmelt-driven freshet was occurring 10 days earlier. Najafi *et al.* (2017) reported declines in spring (1 April) snow water equivalent (SWE) in the upper Peace, Fraser, and upper Columbia river basins. Similarly, declines in spring snow cover extent were detected (Déry and Brown 2007; Brown and Mote 2009; Choi *et al.* 2010; Hernández-Henríquez *et al.* 2015) with the most vulnerable regions being the Western Cordillera (Brown and Mote 2009; Choi *et al.* 2010) and at low- to mid-elevations (Brown and Mote 2009; Hernández-Henríquez *et al.* 2015).

The integrity of the snowpack is vulnerable to extreme winter weather. In particular, anomalously cold or warm conditions and precipitation phase can affect the structure of the snowpack and have been linked to the generation of hydrologic extremes (e.g. Doyle and Costerton 1993). The frequency and magnitude of winter (DJFM) temperatures above freezing and rainfall in western Canada from 1946-2012 were evaluated and presented in Chapter 2. Both above-freezing temperatures and rainfall increased over the study period, particularly during January and March, although the magnitude of trends varied considerably over the study area.

Large-scale atmospheric circulation is responsible for the movement and distribution of water and energy (Trenberth and Stepaniak 2003), which is reflected in the climatic variability in western Canada. Specifically, the mid-troposphere is characterized by a series of mid-latitude troughs and ridges resulting in meridional flow, or, in the absence of troughs and ridges, zonal flow (Holton 1979). These patterns of airflow direct the movement of surface high- and low-pressure systems and the movement of warm or cold, moist or dry air masses.

Numerous studies have evaluated links between atmospheric circulation patterns and surface climate and hydrology. A mid-tropospheric ridge of high pressure centred over western Canada has been linked to above-average temperatures and below-average precipitation while a ridge of high pressure centred over the Pacific Ocean and adjacent trough over western Canada is associated with below-average temperatures and above-average precipitation in western Canada (Romolo *et al.* 2006a,b; Newton *et al.* 2014a; Bonsal *et al.* 2017; Bonsal and Cuell 2017). Romolo *et al.* (2006a) determined that winter snow accumulation in the Peace River Basin increased with a higher frequency of zonal flow or a trough of low pressure over western Canada compared with a ridge of high pressure over western Canada. In the same region, Romolo *et al.* (2006b) found that a mid-tropospheric ridge of high pressure over western Canada is linked to the onset of spring snowmelt. Newton *et al.* (2014a) determined that a strong ridge of high pressure in the mid-troposphere, whether centred over the Pacific Ocean or western Canada, exhibited strong persistence, frequently occurring over multiple consecutive days.

The persistent meridional flow associated with high-amplitude ridges and troughs is linked to extreme weather in North America (Francis and Vavrus 2012; Petoukhov *et al.* 2013; Screen and Simmonds 2014). Newton *et al.* (2017) found that a persistent ridge of high pressure over western Canada was a contributing driver of numerous mid-winter river ice break-up events. Fitzharris (1987) described patterns of surface high- and low-pressure systems as they relate to major avalanche winters in southwestern British Columbia and determined that persistent cold Arctic outbreaks followed by warm, Pacific frontal systems were conducive to major avalanche activity, highlighting the sequencing of large-scale circulation for the generation of extreme events. Hydrological responses may not be a linear function of climatic variability, but rather the product of a threshold exceedance or tipping point (McGregor 2017), emphasizing the importance of understanding links between persistence and extreme weather and hydrologic phenomena.

Interannual variability of climate in western Canada has been linked to large-scale teleconnection patterns that act on interannual and interdecadal time scales, including the El Niño-Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO). However, the surface climate responses to positive and negative phases of ENSO and the PDO were found to be a function of the influence of these teleconnections on the frequency of dominant atmospheric circulation patterns (Romolo *et al.* 2006a,b; Stahl *et al.* 2006; Newton *et al.* 2014a). Previously, El Niño (negative Southern Oscillation Index; SOI) and positive phases of the PDO were associated with above-average winter temperatures and below-average precipitation in western Canada, while La Niña (positive SOI) and negative phases of the PDO are associated with below average temperatures and above-average precipitation (Shabbar and Khandekar 1996; Shabbar *et al.* 1997; Bonsal *et al.* 2001). Recently, Newton *et al.* (2014a) described an increase in the frequency of a ridge of high pressure over western Canada during El Niño and positive phases of the PDO, particularly when these two teleconnection patterns coincide. Conversely, a ridge of high pressure over the Pacific Ocean and adjacent trough of low pressure over western Canada, as well as zonal flow over western North America, occurred with a higher frequency during La Niña and negative phases of the PDO (Newton *et al.* 2014a).

Although several studies have examined relationships between surface climate and mid-tropospheric circulation patterns or atmospheric or oceanic teleconnections, none has examined the role of atmospheric circulation on the frequency and magnitude of winter above-freezing temperatures and rainfall. Given the high risk of diminishing snowpack to water security in western Canada and the potential for the generation of hydrologic extremes, it is valuable to improve our understanding of large-scale atmospheric drivers of winter climate variability. Therefore, this research identifies the synoptic-scale mid-tropospheric circulation patterns associated with temperature and precipitation patterns conducive to snowmelt or degradation of the snowpack during the winter season in western Canada. Specifically, dominant atmospheric circulation patterns in the mid-troposphere are identified and the corresponding frequency and magnitude of above-freezing temperatures and associated rainfall are calculated.

3.2. Study Area

This research is focused on major river basins in western Canada, spanning varied hydroclimatic and physiographic regions including the Western Cordillera, boreal forest, and Prairies (Figure 3-1). The Liard River flows from alpine headwaters through boreal regions in northeastern BC and southeastern Northwest Territories and is a major tributary of the Mackenzie River. Similarly, the Peace and Athabasca Rivers flow from mountain headwaters, across the northern Prairie and Boreal forest regions to the Peace-Athabasca Delta, and are also tributaries of the Mackenzie River. The Saskatchewan River flows east from the Rocky Mountain headwaters and across the southern Prairies, ultimately contributing to the Nelson River and draining into Hudson Bay. The Stikine, Nass, and Skeena rivers are located on the north coast of BC and drain into the Pacific Ocean. The Fraser and Columbia rivers originate on the western slopes of the Rocky Mountains and eastern slopes of the Columbia Mountains and drain into the Pacific Ocean. These rivers are snowmelt-dominated, with peak flows occurring in late spring or early summer, coinciding with snowmelt. Summer streamflow is a function of rainfall and glacier melt. Flows decrease in the autumn and remain low throughout the winter, with many rivers forming a solid ice cover.

The climate of western Canada is strongly influenced by warm, moist air masses originating over the Pacific Ocean and cold, dry air masses originating over the Arctic Ocean and northern Canada. Precipitation is highest along the coast and decreases with increasing distance from the Pacific Ocean. The convergence of moist Pacific and cold Arctic air masses can result in heavy, dense snowfall, particularly over BC (Geng *et al.* 2012). Atmospheric rivers, originating over the sub-tropical Pacific Ocean, are infrequent, but have the potential to deliver a concentrated band of moisture and heat over western Canada (Roberge *et al.* 2009). The windward slopes of mountain regions receive higher precipitation as moist air masses are forced to rise and release moisture, while leeward regions such as the Fraser Plateau and the Prairies receive lower precipitation. Chinook winds, dry adiabatically warmed air masses that descend the leeward side of mountains, frequently occur to the east of the Rocky Mountains, particularly in southern Alberta (Longley 1967; Goulding 1978).

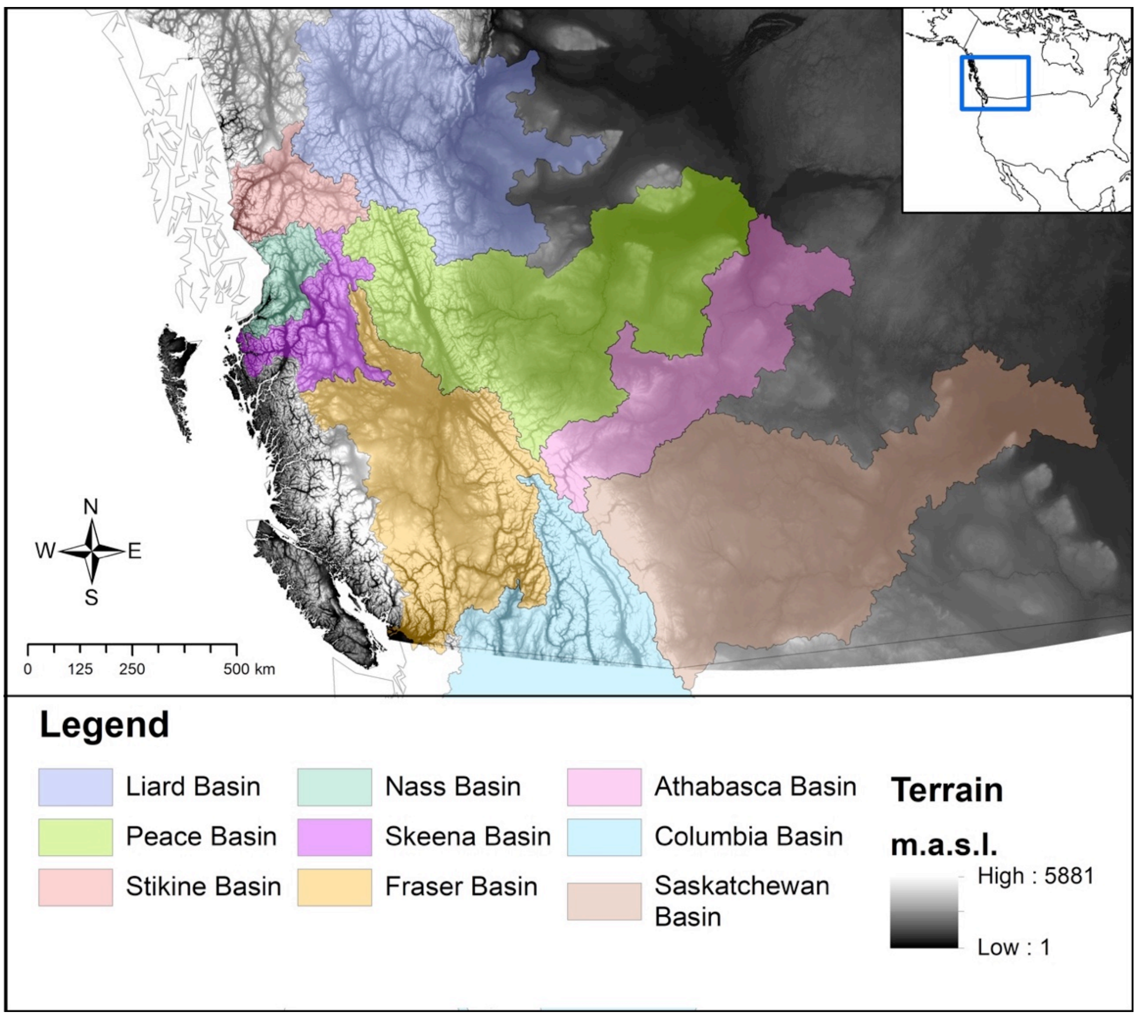

Figure 3-1: Major river basins in western Canada that are the focus of this research

3.3. Data and Methods

Daily winter (DJFM) geopotential height (GPH) data at 500 hPa, between 30°N and 70°N and 100°W and 170°W, from 1949-2012, obtained from the National Centers for Environmental Prediction/National Centre for Atmospheric Research (NCEP/NCAR; Kalnay *et al.* 1996), are classified into a set of dominant patterns using the Self-Organizing Maps (SOM) Toolbox for Matlab (Vesanto *et al.* 2000). This synoptic window captures atmospheric flow pathways from the Pacific and Arctic Oceans as well as circulation features over western North America. SOM is public domain software, available from the Laboratory of Computer and Information Science Adaptive Informatics Research Centre at Aalto University in Epsoo, Finland (http://www.cis.hut.fi/research/som-research/). SOM is a statistical tool in the field of artificial neural networks that clusters data and arrange them onto a topologically ordered array such that spatial and temporal relationships between daily patterns are preserved (Kohonen 2001). The atmospheric circulation pattern for a particular day is related to the previous and next day, and these relationships are captured through SOM classification. Maximum variance exists in the opposite corners of the array while neighbouring patterns are most similar. Thus, SOM presents an advantage over alternative methods of classification as atmospheric circulation patterns are not discrete, but rather evolve over time. The organizational capabilities of SOM give rise to a visual representation of atmospheric states that facilitates analyses among synoptic types and with surface climate variables. Mathematical distances between synoptic type vectors on the SOM array are commonly calculated and visualized using a Sammon map (Sammon 1969). Although the SOM array demonstrates visual similarities among patterns, the Sammon map quantifies similarities between data vectors, facilitating the grouping of similar patterns to evaluate persistence of synoptic regimes. A comprehensive description of SOM is found in Kohonen (2001) and SOM applications to synoptic circulation classification can be found in Hewitson and Crane (2002), Reusch *et al.* (2005), Reusch (2010), Sheridan and Lee (2011), and Newton *et al.* (2014a,b).

A number of metrics are used to statistically describe the relationships between and temporal evolution of atmospheric states. The frequency of each synoptic type is calculated for each winter season and trends are evaluated using the Mann-Kendall non-parametric test (MK; Mann 1945; Kendall 1975), using significance levels of $p < 0.05$ and $p < 0.10$, and Sen's method for slope estimation is used to determine the magnitude of the trends (Sen 1968). Synoptic type frequencies were evaluated for change points to determine if there was an abrupt shift in the time series. Three statistics were selected to evaluate change points: mean, standard deviation, and slope. For the distribution of each synoptic type, the point at which the statistical metric changes most abruptly is identified using change point analysis in the Matlab programming platform. The time series for each synoptic type is then divided into two segments based on the change point identified for each metric and compared using the two-sample non-parametric Kolmogorov-Smirnov (KS) test to evaluate whether the two series are from the same continuous distribution.

High-resolution (1/16-degree), gridded daily winter (DJFM) minimum and maximum air temperature (°C) and precipitation (mm) data for western Canada, from 1949 to 2012, are used to evaluate surface climate variables associated with synoptic types. The dataset was developed using a thin-plate spline interpolation of climate station data, using ClimateWNA climatology (Wang *et al.* 2012) as a covariate (Werner *et al.* 2018). Daily maximum and minimum temperatures are averaged to estimate mean daily temperature. Days when the mean daily temperature is above freezing (T_{mean} > 0°C) are identified and accumulated melting degree-days (MDD) are calculated as the sum of mean daily temperatures above freezing throughout the winter season. Rainfall is identified using a temperature-index precipitation phase equation, whereby precipitation falling on days when the mean daily temperature is equal to or above 1°C is considered rain, and below 1°C is snow, as used in previous studies (USACE 1956; Rohrer 1989; L'hôte *et al.* 2005; Yuter *et al.* 2006; Lundquist *et al.* 2008; Kienzle 2008). The greatest uncertainty for precipitation phase determination exists between 0°C and 2°C (Feiccabrino *et al.* 2012) and at temperatures nearing 0°C precipitation may be mixed rain and snow, slush, graupel, or hail; however, these precipitation types may be associated with sufficient heat energy to generate snowmelt (USACE 1956). The percentage of days when the mean daily temperature is above freezing and rainfall occurs are calculated for each identified synoptic type.

3.4. Results

Mid-tropospheric circulation

Daily winter 500 hPa GPH were classified into 12 types on a topologically organized 3 × 4 array using SOM, which is large enough to identify dominant atmospheric circulation patterns and small enough to capture differences between patterns (Figure 3-2). The synoptic types in the SOM array are numbered according to the position on the array. Atmospheric flow direction in the mid-troposphere is roughly parallel to the contour lines and directs surface high- and low-pressure systems (Holton 1979). Types 1 and 4 in the top left corner of the SOM array are characterized by a strong ridge of high pressure extending over the Pacific Ocean and Alaska and adjacent trough of low pressure over western Canada, indicative of northerly meridional advection of cold Arctic air over western Canada. Conversely, a ridge of high pressure over western Canada (Types 7-12) directs warm Pacific air masses toward coastal BC and blocks the movement of cold, Arctic air masses from entering the region. These circulation types have been previously linked to anomalously warm, dry surface climate, where the magnitude of the surface climate response is related to the strength and position of the ridge (Bonsal *et al.* 2001; Stahl *et al.* 2006; Newton *et al*. 2014a). Zonal flow patterns (Types 2, 3, 5, and 6) indicate a lack of surface high- and low-pressure systems and unobstructed airflow from the Pacific Ocean over the study region. Zonal flow during the winter season is associated with above-average precipitation and below-average temperatures (Romolo *et al.* 2006a,b; Newton *et al.* 2014a).

Synoptic type frequency, persistence, and trajectory together describe the evolution and dominant states of atmospheric circulation. Synoptic types in the four corners of the SOM array, Types 1, 3, 10, and 12, occur with the greatest frequency (Figure 3-3a), while intermediate types, particularly in the centre of the SOM array (Types 5 and 8) are infrequent transition patterns, facilitating the shift from one dominant atmospheric state to another. Types characterized by a strong ridge of high pressure, Types 1, 10, and 12, are the most persistent, with an average persistence of 76%, 70%, and 73%, respectively (Figure 3-3b). Type 1 persists for an average of four days, but has persisted for up to 34 consecutive days. Similarly, Types 10 and 12 persist for an average of three and four days and up to 19 days and 28 days, respectively. Extreme weather phenomena have been linked to persistent atmospheric circulation patterns, particularly those patterns characterized by strong meridional flow, such as Types 1, 10, and 12 (Francis and Vavrus 2012; Petoukhov *et al.* 2013; Screen and Simmonds 2014). Zonal flow (Type 3) is also highly persistent, with an average persistence of 67%, occurring for an average of three days, and persisting for up to 14 consecutive days. Trajectory (Figure 3-3c) indicates preferred shifts from one synoptic type to neighbouring patterns, where the length of the arrow is proportional to the frequency of shifts from one pattern to another. It is evident that the preferred trajectory follows the outer patterns along the array with approximately equal frequency in either direction.

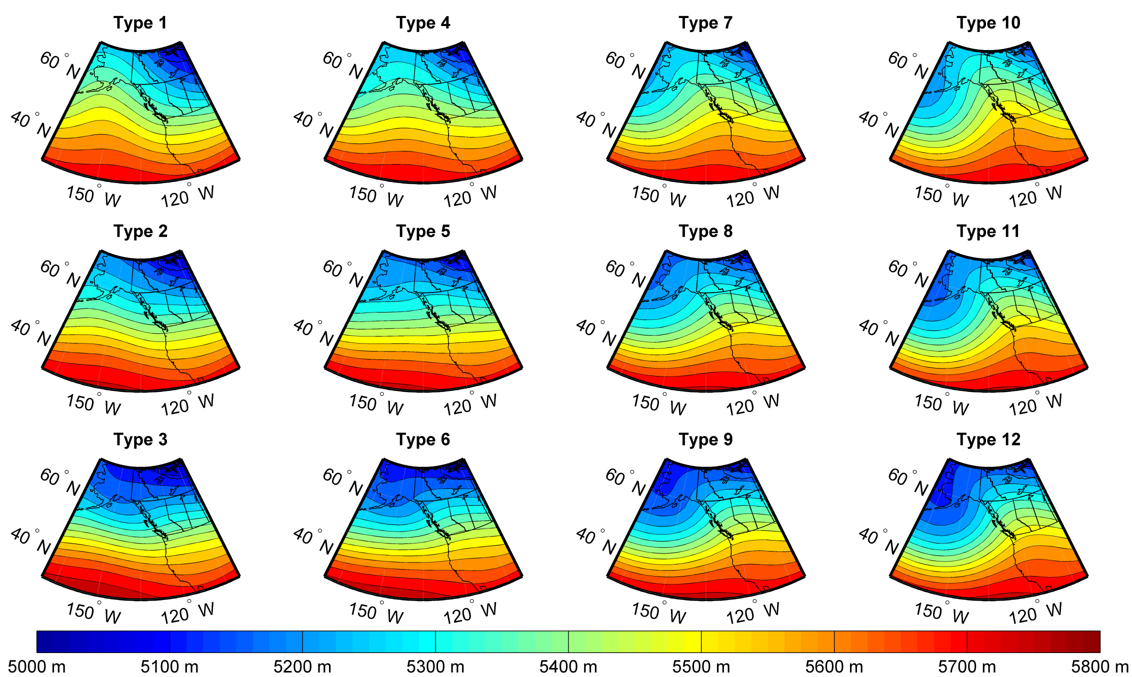

Figure 3-2: Daily winter (DJFM) geopotential heights at 500 hPa from 1949-2012, classified using Self-Organizing Maps

a)

12.8	6.6	6.8	10.6
6.2	4.5	4.5	6.5
10.8	9.3	8.9	12.5

b)

76	47	46	70
42	28	30	48
67	47	48	73

c)

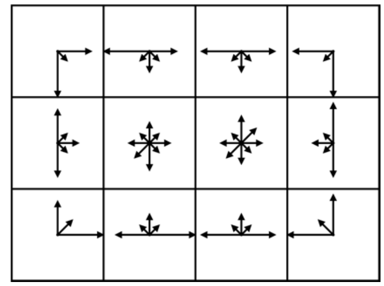

Figure 3-3: Synoptic type a) frequency (%), calculated as a percentage of total days in the study period, b) persistence (%), calculated as the percentage of occurrences where the pattern remains the same on the following day, and c) trajectory indicating preferred shifts in atmospheric states, where the length of the arrow is proportional to the frequency of shifts, from 1949-2012

In general, the frequency of synoptic types is fairly evenly distributed throughout the winter season (Figure 3-4). Types 9 and 12 are slightly more frequent during Dec-Feb compared with March, while Type 10 is more frequent from Jan-Mar than December. Interannual variability of synoptic type frequency is high, particularly for the four dominant patterns, Types 1, 3, 10, and 12 (Figure 3-5). For example, the frequency of Type 1 is as high as 36% of winter days, while Type 12 is up to 50%. Significant decreases in synoptic type frequency are seen in Type 1 ($p < 0.05$) and Type 3 ($p < 0.10$), and Type 10 has significantly increased ($p < 0.05$) over the study period, suggesting a decrease in both high-pressure ridging over the Pacific Ocean and zonal flow, and an increase in ridging over British Columbia. Bonsal *et al.* (2001) and Newton *et al.* (2014a) found that a ridge of high pressure over western Canada dominated winters categorized by positive phases of the PDO and negative phases of the SOI (El Niño), particularly when positive PDO and El Niño occurred simultaneously. There are several high-frequency peaks evident in the second half of the time series of Type 12 (Figure 3-5), the two most frequent of which (50% in 1983 and 37% in 1998), coincide with a strong negative SOI and positive PDO (Mantua and Hare 2006; Shabbar 2006).

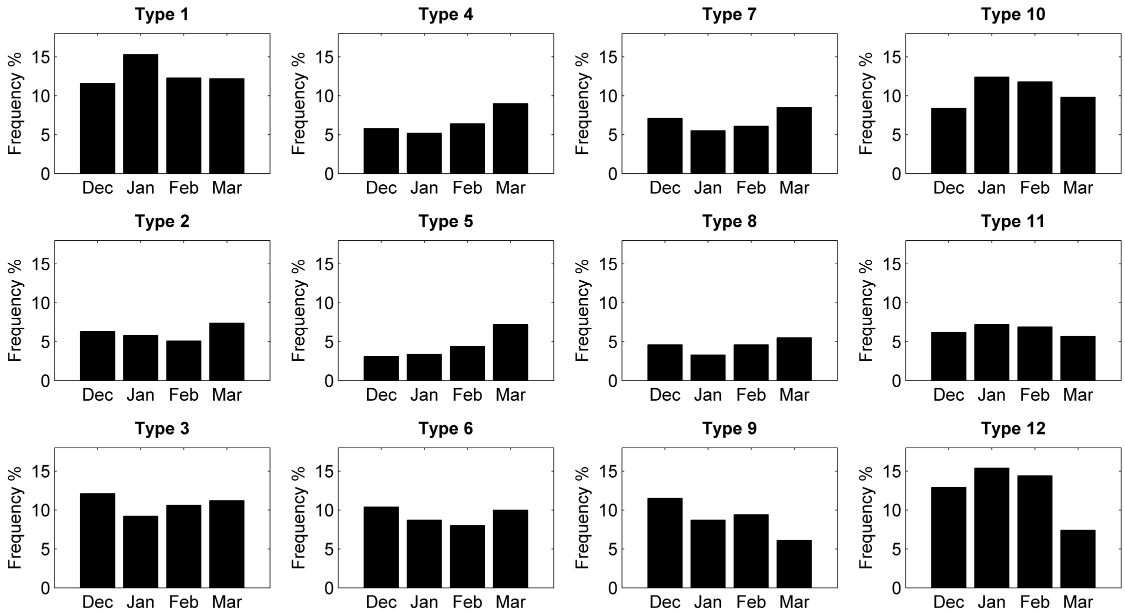

Figure 3-4: Average monthly synoptic type frequency (%) for the period 1949-2012

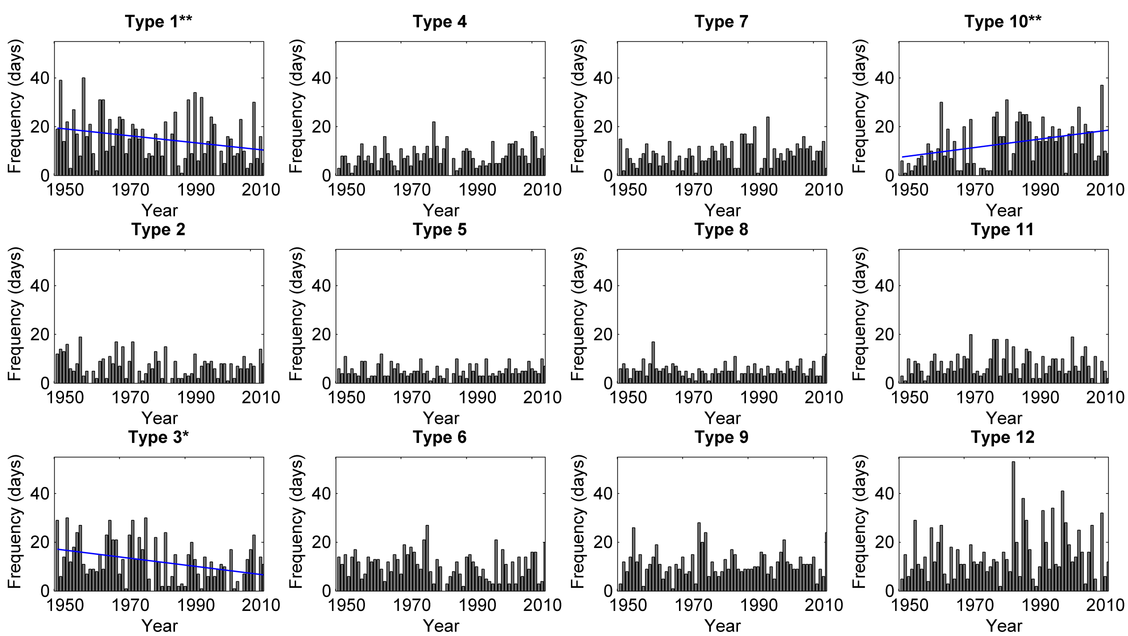

Figure 3-5: Time series of synoptic types from 1949-2012. Trends, evaluated using the Mann-Kendall non-parametric test for trend. Only those trends significant at 10% (denoted by *) and 5% (denoted by **) are shown (blue lines)

Types 3 and 10 have a change point in 1977 for all three metrics, and results of the KS test indicates that for both Type 3 and Type 10 the two distributions, 1949-1976 and 1977-2012 are significantly different ($p < 0.05$). Despite the appearance of a change in frequency and variability of Type 12, the analysis failed to detect a change point that divided the time series into two significantly different distributions. The mean frequency of Type 3 was higher prior to 1977 than after, and the slope suggests a linear increase in both time series (Figure 3-6a), contrary to the decreasing linear trend detected by the MK test. The mean frequency of the 1949-1977 time series of Type 10 is lower than 1977-2012 and the slope is minimal prior to 1976 and decreasing after 1977, contrary to the increasing linear trend detected by the MK test (Figure 3-6b). Additionally, the average seasonal persistence of synoptic types, measured as the percentage of days each winter when that type occurs for consecutive days, was evaluated for trend and change points. A significant increasing trend and a step-change increase in 1977 in the mean persistence were detected for Type 10 (Figure 3-7). These step-changes coincide with a documented shift from a predominantly negative to positive phase of the PDO (Mantua *et al.* 1977; Mantua and Hare 2002), which has been linked to anomalous surface climate and streamflow in western Canada, including links between positive phases of the PDO and positive winter temperature anomalies (Bonsal *et al.* 2001), lower precipitation, particularly in coastal regions (Fleming and Whitfield 2010), and lower streamflow (Mantua *et al.* 1997; Déry and Wood 2005), with opposite hydroclimatic impacts during negative phases of the PDO.

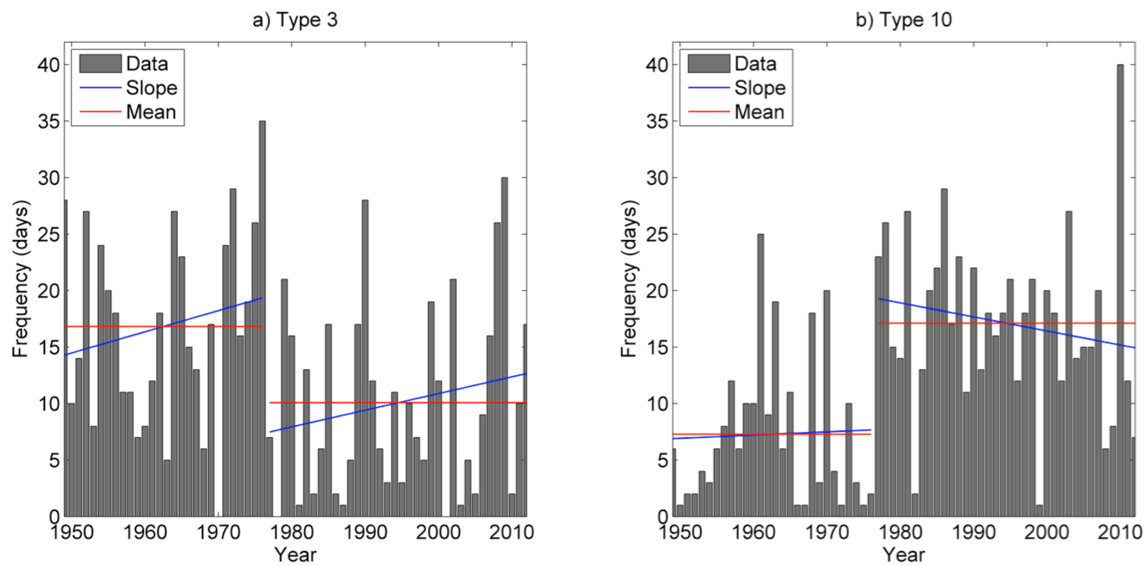

Figure 3-6: Step-changes in the mean and slope of the frequency of a) Type 3 and b) Type 10, from 1949-2012

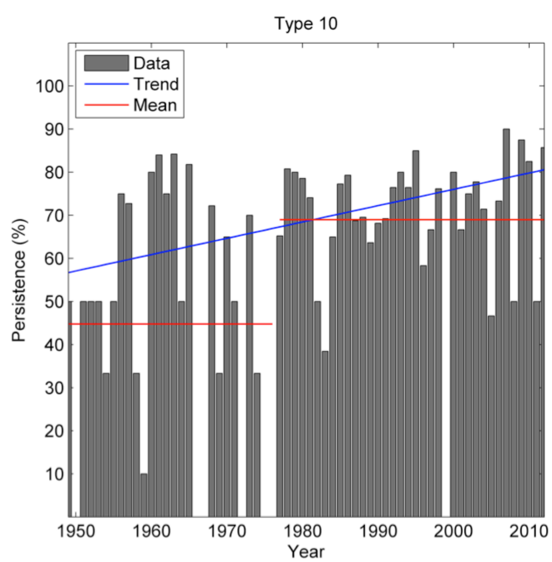

Figure 3-7: Persistence of Type 10, where persistence is defined as the percentage of days when Type 10 occurs the following day, from 1949-2012. Red lines indicate mean persistence before and after the shift in 1977 and the blue line indicates trend, significant at $p < 0.05$

Surface above-freezing air temperature and associated rainfall

The frequency of days when the mean daily temperature is above freezing for each synoptic type was calculated as a percentage of the total distribution at each grid point for that type. The position of each pattern of above-freezing surface air temperature corresponds to the synoptic type in the same position on the SOM array (Figure 3-8). The frequency of above-freezing temperatures increases gradually from the upper left corner (Type 1) to the lower right corner (Type 12) of the array. The frequency of above-freezing temperatures is high across much of the study region during days when there is a ridge of high pressure over western Canada, and the strength of the surface climate response is dependent on the strength and position of the ridge (Types 9-12). For example, the frequency of above-freezing temperatures associated with Type 12 approaches 100% along the coastal region, 50% in the low-elevation regions of the Fraser and Columbia basins and the upper Saskatchewan basin, and 40% in portions of the upper Peace and upper Athabasca basins. Above-freezing temperatures associated with Types 8, 9, and 11 are similar to Type 12, albeit with a lower frequency. Conversely, Type 1 in the opposite corner of the array is associated with very few days when the mean daily temperature is above freezing, with up to 60% of winter days in near-shore coastal areas and up to 20% of winter days in the Fraser, Columbia, and north-coastal basins above freezing. Type 3 exhibits a low frequency of above-freezing temperatures, primarily seen in the southern half of the study region. These patterns of above-freezing temperatures are consistent with negative temperature anomalies associated with a ridge of high pressure over the Pacific Ocean and adjacent trough of low pressure over western Canada, directing the flow of cold, dry air masses from the Arctic to western Canada and westerly zonal flow from the Pacific Ocean over the study region, and the positive temperature anomalies associated with a ridge of high pressure over western Canada (Bonsal *et al.* 2001; Romolo *et al.* 2006b; Newton *et al.* 2014a).

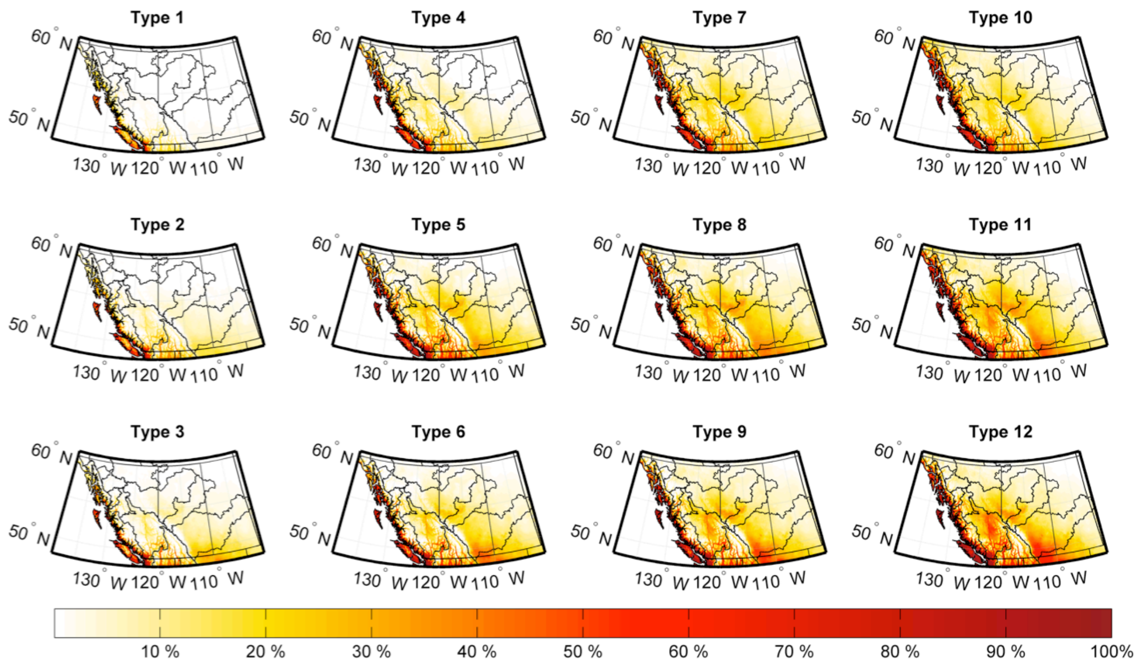

Figure 3-8: Percentage of winter days when the mean daily temperature is above freezing for each synoptic type, from 1949-2012

Type 10 increased over the study period, while Types 1 and 3 have decreased, suggesting a reduction in the frequency of mid-tropospheric circulation conducive to colder temperatures and an increase in circulation conducive to above-freezing temperatures. Additionally, the persistence of Type 10 has increased over the study period, indicating longer periods of above-freezing temperatures. This provides an atmospheric mechanism for increasing trends in the frequency and magnitude of above-freezing temperatures from 1946-2012 reported in Chapter 2. Furthermore, Types 10 and 12 exhibit high persistence, and Newton *et al.* (2017) found that a prolonged, persistent ridge of high pressure over western Canada and associated surface hydroclimatic variables preceded numerous mid-winter river ice break-up events.

Average daily precipitation was calculated for each synoptic type (Figure 3-9). Synoptic types located on the top row of the SOM array (Types 1, 4, 7, and 10) are associated with low precipitation compared with those types along the bottom row (Types 3, 6, 9, and 12). Specifically, Type 1 is associated with low precipitation across the study region with slightly higher precipitation (up to 5 mm) along the coast and Type 10 is associated with higher precipitation along the coast, but very low precipitation inland. Type 10 is characterized by a strong ridge of high pressure with a ridge axis centred near the coast, which effectively blocks moisture inflow to the study region. Days with zonal flow (Type 3) see higher precipitation along the coast and the Rocky Mountains with low-moderate precipitation in the remainder of the study region. Type 12, a ridge of high pressure over western Canada, is associated with high precipitation along the coast and low to no precipitation east of the Rocky Mountains.

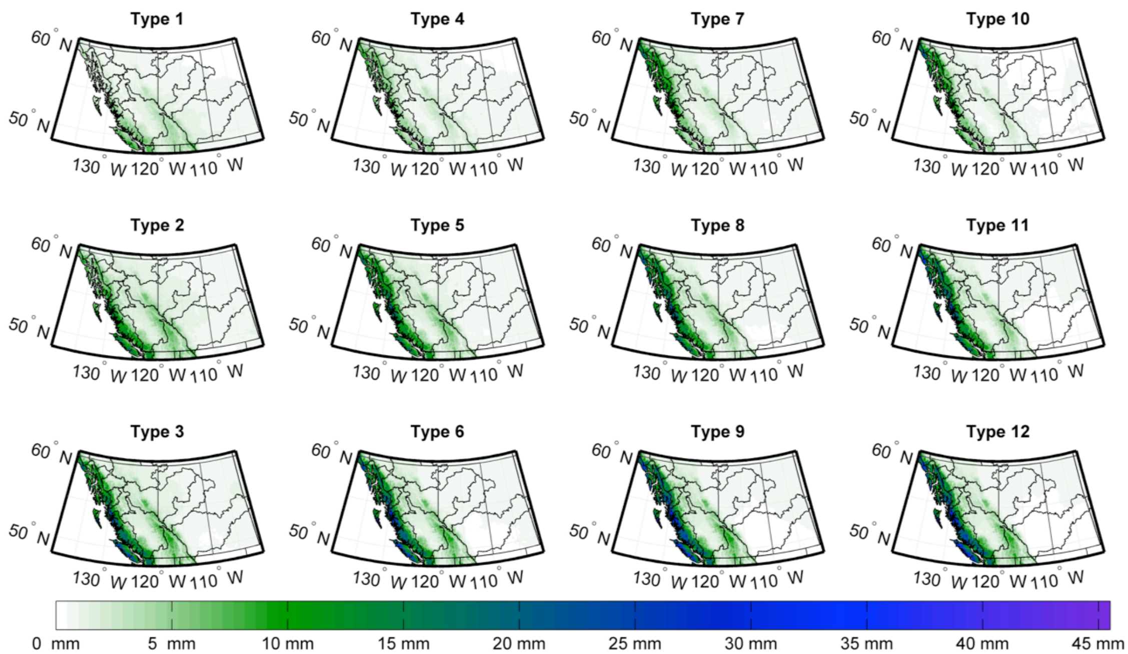

Figure 3-9: Average precipitation associated with each synoptic type, from 1949-2012

The percentage of days when rainfall occurred for each grid point was calculated for each synoptic type (Figure 3-10). A very low frequency of rainfall, confined to the southern coastal region of the study area, is associated with Types 1, 2 and 3. A moderate to high frequency of rainfall (> 50%) is seen along the coast and low (< 30%), but widespread rainfall is found in the Columbia, Fraser, upper Peace, and north coastal basins during days classified as Type 12. Similar spatial patterns of rainfall frequency, at a smaller magnitude, are associated with Types 9 and 11. These synoptic types are characterized by a ridge of high pressure over western Canada, which effectively blocks the advection of moisture into the study region, particularly east of the Rocky Mountains; however, these types are associated with a high frequency of above-freezing temperatures increasing the likelihood of precipitation falling as rain. Type 10 is also characterized by a blocking ridge of high pressure over western Canada, with a ridge axis centred near the coast, and is associated with lower rainfall across the study region compared with Types 9, 11 and 12. Types 5 and 8 are infrequent transition patterns, but are associated with a low (< 20%) frequency of rainfall across the Saskatchewan and Athabasca river basins. Zonal flow (Type 3) is associated with a very low frequency of rainfall, except along the coast. Zonal flow is conducive to moisture advection from the Pacific Ocean over the study region; however, it is also associated with a relatively low frequency of above-freezing temperatures. Thus, the precipitation seen with Type 3 falls primarily as snow. These results are consistent with results presented in Chapter 2 showing that winter rainfall was more prolific in the mountainous Western Cordillera and less frequent east of the Rocky Mountains. Similarly, Newton *et al.* (2017) found that the majority of mid-winter river ice break-up events in British Columbia were triggered by rain-on-snow while events in the region east of the Rocky Mountains, were commonly triggered by periods of above-freezing temperatures.

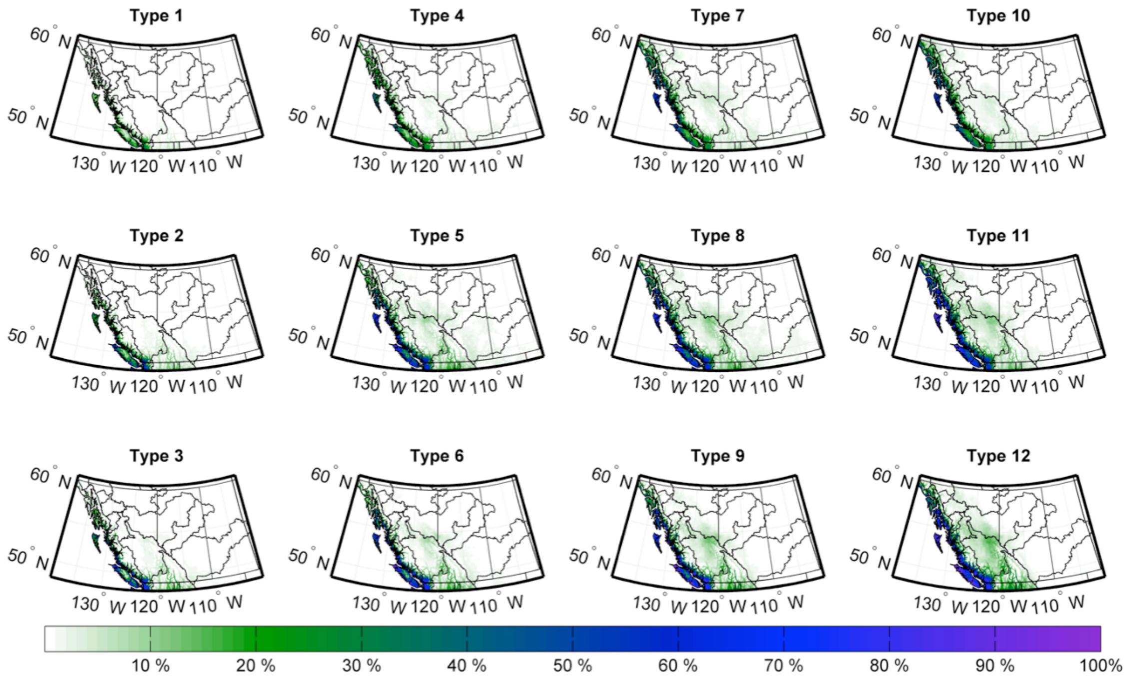

Figure 3-10: Percentage of days with rainfall for each synoptic type, from 1949-2012

High frequency, persistent circulation

Relative spatial distances between synoptic types on the SOM array can be calculated and visualized using Sammon maps (Sammon 1969). Similar patterns cluster closer together on the Sammon map, while dissimilar patterns lie further apart. The Sammon map (Figure 3-11) reveals a divide whereby the six synoptic types on the left hand side of the array are similar while the six types on the right side are similar. The individual synoptic types are grouped into quadrants to facilitate analysis of synoptic regime persistence (Figure 3-12). Types 8, 9, 11 and 12, depicting a ridge of high pressure over western Canada, are grouped in the lower right (LR) quadrant of the SOM array. Types 7 and 10, in the upper right (UR) quadrant, are also indicative of a ridge of high pressure, but the ridge axis is centred over the coastal region and extends further north. In the upper left (UL) quadrant are Types 1 and 4, characterized by a ridge of high pressure over the Pacific Ocean and adjacent trough over western Canada. Zonal flow patterns, Types 2, 3, 5 and 6, are located in the lower left quadrant (LL).

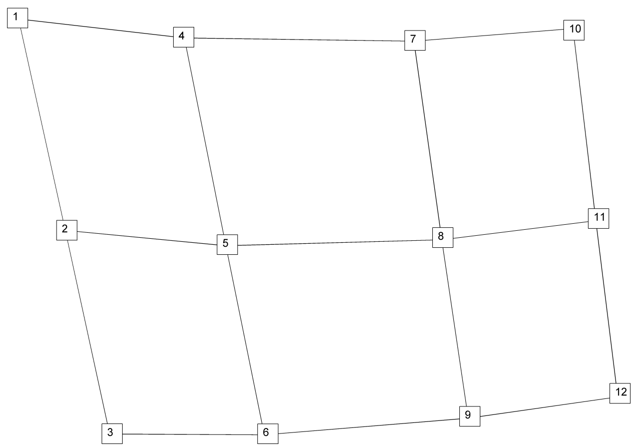

Figure 3-11: Sammon map for the SOM array

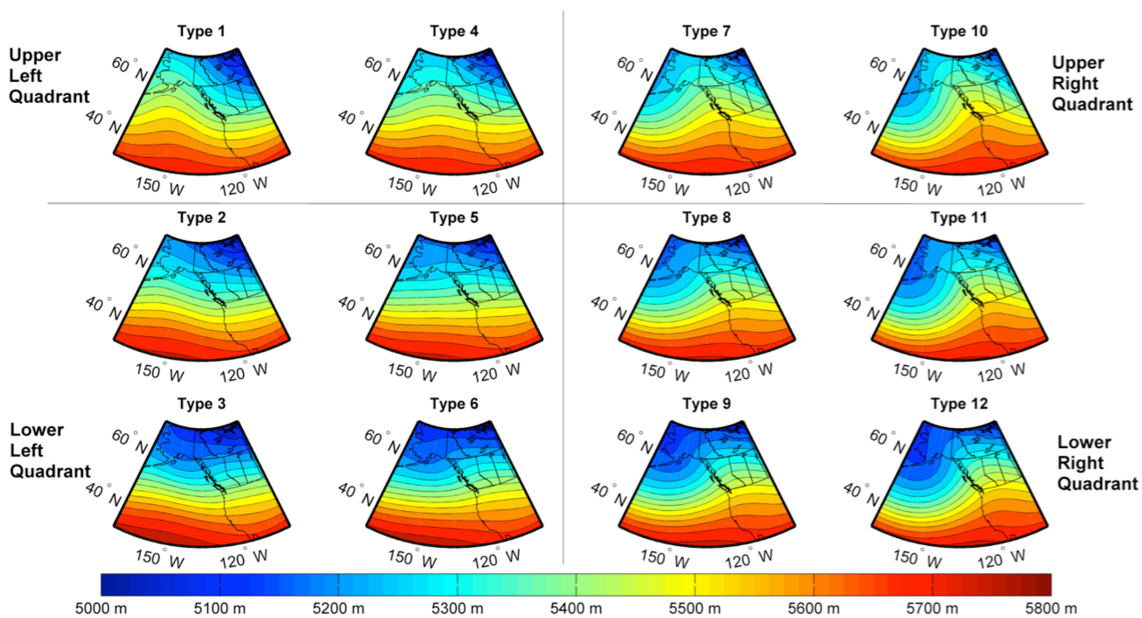

Figure 3-12: SOM divided into four quadrants depicting four dominant synoptic regimes, a ridge of high pressure over the Pacific Ocean (UL), zonal flow (LL), a ridge of high pressure centred near the coast (UR), and a ridge of high pressure over western Canada (LR)

The average winter frequencies of the UL, UR, LL, and LR regimes are 19.4%, 17.4%, 30.8%, and 32.4%, respectively; however, there are numerous years when one of these regimes is dominant and highly persistent. For example, in 1983, synoptic types in the LR quadrant (Types 8, 9, 11, and 12) occurred 73.6% of winter days, and persisted up to 40 consecutive days. The persistence of synoptic type regimes was evaluated for trend and change-points. A significant ($p < 0.05$) increasing trend was detected for the upper right corner (UR) as well as a step-change in the mean persistence in 1977 (Figure 3-13). These results are consistent with a trend and step-change identified for Type 10. A step-change decrease was detected for the synoptic regime in the upper left corner (UL), in 1981, and for the regime in the lower left corner, in 1978. The step-change and linear increase in UR signifies an increase in the potential for longer periods of above-freezing temperatures and rainfall.

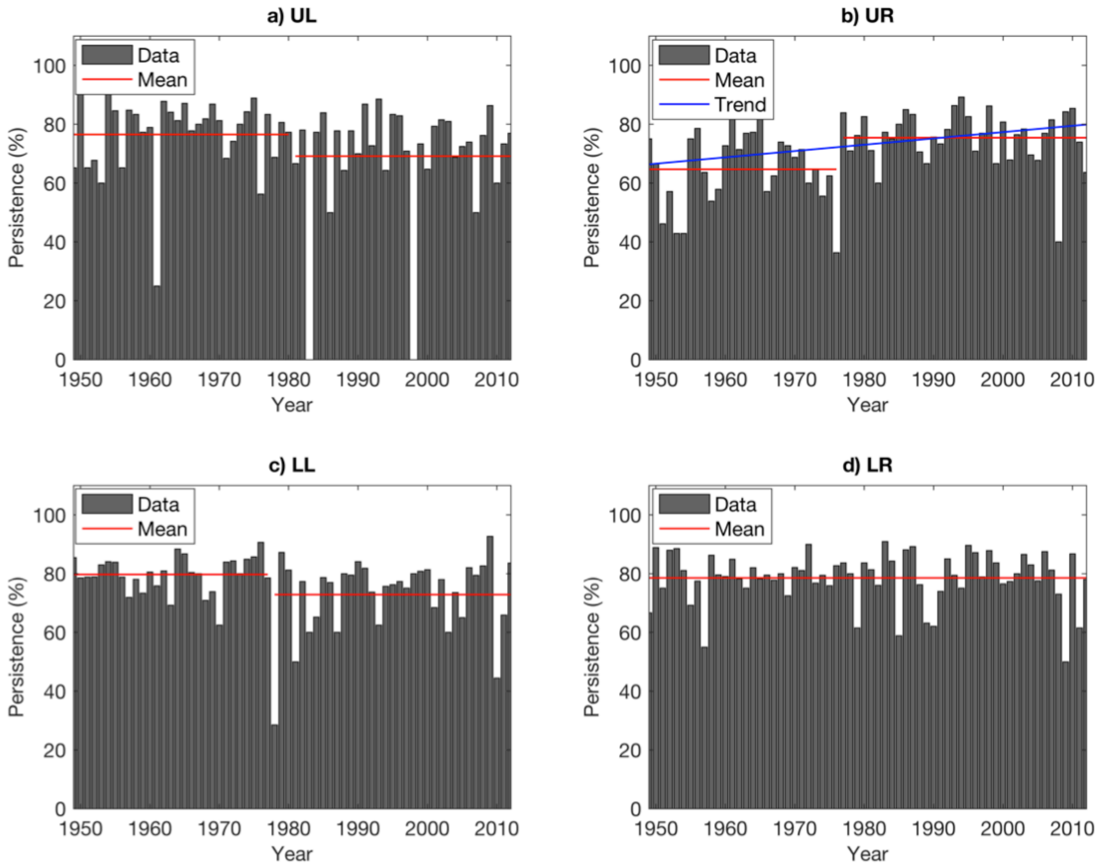

Figure 3-13: Persistence of synoptic regimes, where persistence is calculated as the percentage of days when the same synoptic regime occurs the following day, from 1949-2012. Red lines indicate mean persistence before and after a step-change (UL in 1981, UR in 1977, LL in 1978, and LR no step-change), and blue line indicates trend, significant at $p < 0.05$

To capture the surface climate response to the four synoptic regimes over western Canada, winter accumulated MDD and total rainfall anomalies were calculated for winters when the frequency of each quadrant was in the top 10% of each regime frequency distribution. Several years with strong El Niño events appear in the LR list, including 1983, 1992 and 1998 (Shabbar 2006), which also coincide with the warm, or positive phase of the PDO (Mantua and Hare 2002), indicating a strong influence of these teleconnections on atmospheric circulation persistence. Three winters were strongly dominated by synoptic types on the right half of the SOM array, 1981 (48.8% LR and 31.4% UR), 1986 (48.8% LR and 33.1% UR), and 1987 (53.7% LR and 30.6% UR), while synoptic types in 2009 were predominantly on the left half of the SOM array (46.3% LL and 36.4% UL).

Results reveal negative accumulated MDD anomalies in the left quadrants and positive anomalies in the right quadrants (Figure 3-14). In particular, winters dominated by UR and LR are characterized by accumulated MDD that were up to 150 MDD above normal along the coast and low-elevation river valleys in the Fraser and Columbia river basins, and up to 100 MDD above normal in the upper Saskatchewan, and 50 MDD above normal in the upper Athabasca and upper Peace River basins. Patterns of accumulated MDD anomalies are similar between UR and LR; however, higher positive anomalies are seen further north in UR, corresponding to the northward extension of the high-pressure ridge. Similarly, rainfall is substantially higher than normal when mid-tropospheric circulation is dominated by UR and LR, on the right side of the SOM array, compared to UL and LL (Figure 3-15). Winter rainfall anomalies of up to 100 mm fall in the north-coastal, upper Peace, Fraser, and Columbia river basins, except for coastal portions of these watersheds, with higher rainfall at lower elevations, and further north when UR dominates compared with LR. Conversely, very little rainfall occurs throughout the winter when the UL and LL regimes dominate.

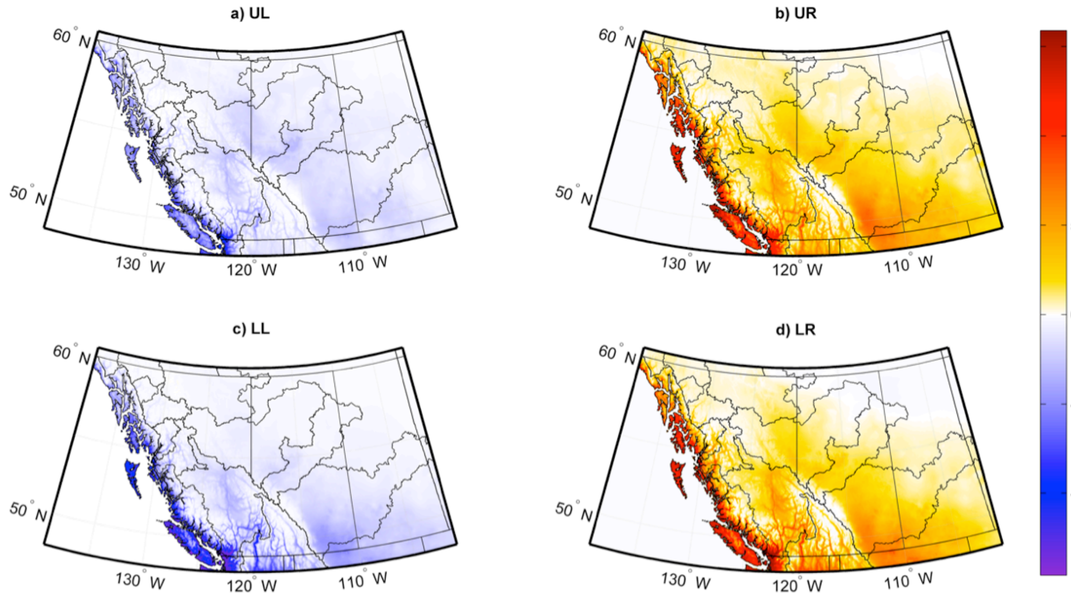

Figure 3-14: Winter accumulated MDD anomalies, relative to the average MDD over 1949-2012, for the four quadrants

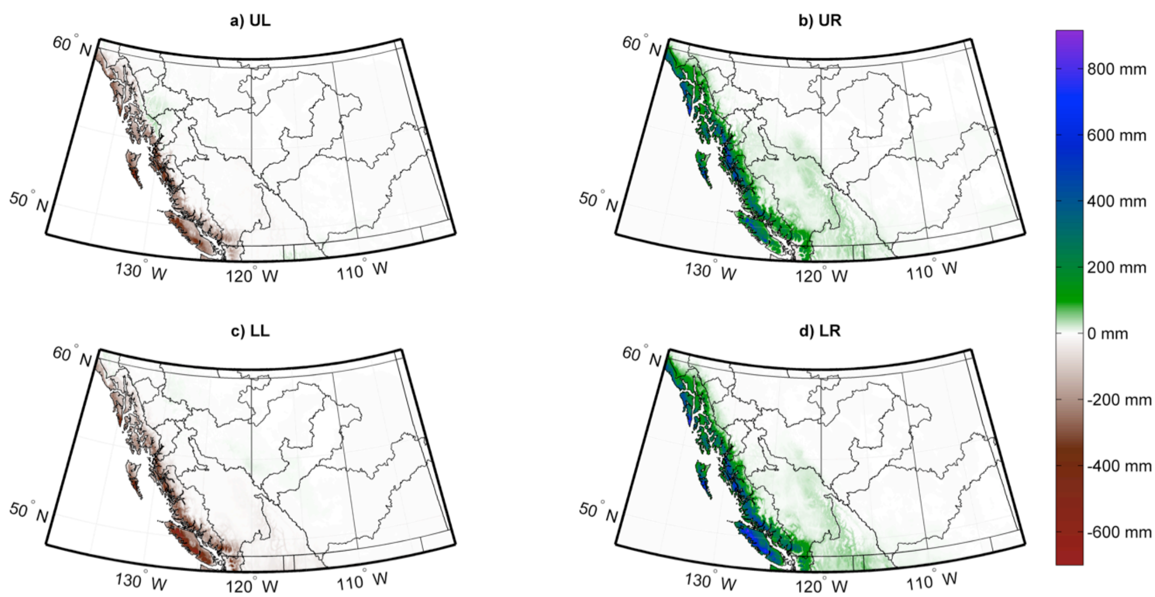

Figure 3-15: Winter rainfall anomalies, relative to the average rainfall over 1949-2012, for the four quadrants

3.5. Discussion and Conclusions

This research has advanced the understanding of the atmospheric drivers of above-freezing winter temperatures and associated rainfall in major river basins in western Canada. The majority of annual streamflow in these rivers originates as winter snowpack and the loss of seasonal snowpack threatens water security and has the potential to generate hydrologic extremes (Sauchyn and Kulshreshtha 2008; Walker and Sydneysmith 2008). Daily winter (DJFM) mid-tropospheric GPH data were classified using SOM to identify dominant circulation patterns and the average frequency of above-freezing temperatures and rainfall was calculated for each synoptic type. The strength of the SOM method of classification is evident from the topological ordering of the resulting SOM array. The organization of the SOM array facilitates the assessment of synoptic type characteristics including frequency, persistence and trajectory, which provide additional insight into the associated surface climate responses. Patterns on the right half of the array depict varying strengths and locations of a ridge of high pressure over western Canada, while patterns in the upper left corner are characterized by a ridge of high pressure over the Pacific Ocean and trough over western Canada, and patterns in the lower left corner are indicative of zonal flow. This enabled the grouping of synoptic types into four regimes to further analyze frequency and persistence.

A ridge of high pressure over western Canada is associated with a high frequency of above-freezing temperatures across the study region and the magnitude of the surface response is related to the strength and position of the ridge. Additionally, the greatest and most spatially widespread frequency of rainfall is associated with a ridge of high pressure over western Canada; however, rainfall is largely confined to the coastal, upper Peace, Fraser, and Columbia river basins, suggesting that these river basins are at the greatest risk of winter rainfall or rain-on-snow, which can contribute both rain and snowmelt volume to runoff (USACE 1956; Colbeck 1975; Male and Gray 1981). Two of the synoptic types with the strongest high-pressure ridging are seen in Type 10, in the upper right corner of the SOM array, and Type 12 in the lower right corner. These two types are frequent, persistent, and elicit a strong surface climate response, suggesting that these types could produce a high volume of winter runoff.

A ridge of high pressure over the Pacific Ocean and adjacent trough of low pressure over western Canada (Types 1 and 4) is associated with a low frequency of above-freezing temperatures and rainfall. This surface response is unsurprising given that this type of circulation directs cold, Arctic air over western Canada and has been associated with negative surface air temperature anomalies in the study region (Romolo *et al.* 2006b; Newton *et al.* 2014a). Similarly, zonal flow (Types 2, 3, 5 and 6) is associated with a low frequency of above-freezing temperatures and rainfall. Previous research identified negative temperature anomalies and positive precipitation anomalies with zonal flow (Romolo *et al.* 2006a,b; Newton *et al.* 2014a). Given the relative low frequency of above-freezing temperatures, precipitation during days with mid-tropospheric zonal flow is likely falling as snow; however, future projections of increasing winter temperatures (e.g., O'Neil *et al.* 2017b; Dibike *et al.* 2017), suggests that zonal flow has the potential to generate more frequent winter rainfall in the future.

The frequency of Type 1, a strong ridge of high pressure over the Pacific Ocean and trough of low pressure over western Canada, significantly decreased over the study period, indicating a decrease of circulation conducive to the suppression of above-freezing temperatures and associated rainfall. Similarly, the frequency of Type 3, characterized by zonal flow, has also significantly decreased over the study period; however, step-change analysis has revealed two linear increases punctuated by a step-change decrease in 1977 resulting in a lower mean frequency. A significantly increasing trend and step-change increase in 1977 were detected in the mean frequency of Type 10. Additionally, the persistence of Type 10 had a step-change increase in 1977. The step-changes in 1977 are consistent with a shift to a positive regime of the PDO (Mantua *et al.* 1997; Mantua and Hare 2002), suggesting additional step-changes may be evident in longer historic and future time series. An increase in the frequency of Type 10 and corresponding surface response, combined with a decrease in Type 1 and Type 3, is consistent with trends in above-freezing temperatures and rainfall revealed in Chapter 2 and decreasing snowpack and earlier snowmelt found in western Canada (O'Neil *et al.* 2017a), the Fraser River Basin (Kang *et al.* 2014, 2016), the Peace River Basin (Romolo *et al.* 2006a,b), and the upper Peace, Fraser, and upper Columbia river basins (Najafi *et al.* 2017).

The presence of step-changes in synoptic-scale mid-tropospheric circulation pattern frequency and persistence is indicative of nonlinear changes in the atmospheric system; however, the timing of the step-changes seen in Types 3 and 10 suggest a relationship with the PDO. Newton *et al.* (2014a) reported linkages between the frequencies of several dominant mid-tropospheric circulation patterns and winters with a strong positive or negative average seasonal PDO index value. This suggests the existence of both linear and nonlinear atmospheric responses to fluctuations in the PDO and the potential for parallel responses in climatic and hydrologic systems. The surface climate associated with mid-tropospheric circulation patterns presented here represent the average conditions related to each synoptic type, but given the step-changes evident in Types 3 and 10, it is likely that there are corresponding, but spatially and/or temporally variable step-changes in temperature and precipitation. McGregor (2017) described the probability of thresholds in atmospheric states required to generate a climatic or hydrologic effect, and results from this research suggest thresholds exist in certain atmospheric states. Further analysis is required to evaluate these thresholds and examine associated climate and hydrologic responses.

The SOM array was divided into four quadrants representing synoptic type regimes. Winters dominated by regimes depicting by various strengths of a ridge of high pressure over western Canada (UR and LR) result in strong positive accumulated MDD and rainfall anomalies. The UR regime exhibits anomalies that extend further north in the study region compared to winters dominated by LR regime. Conversely, the UL- and LL-dominated winters are associated with negative accumulated MDD and rainfall anomalies, suggesting that persistent and/or frequent UL- or LL-type regimes result in fewer days when the mean daily temperature is above freezing and when rainfall occurs. The persistence of UR has significantly increased over the study period, with a step-change increase in 1977. Persistence of meridional atmospheric circulation is associated with extreme weather (e.g., Francis and Vavrus 2012; Petoukhov *et al.* 2013; Screen and Simmonds 2014) and hydrologic phenomena (e.g., Newton *et al.* 2017). An increase in the persistence of a ridge of high pressure over western Canada signifies an increased potential for the generation of snowmelt. UR- and LR-dominated winters are expected to have a thinner snowpack and a higher probability of an extreme hydrologic event. A thinner snowpack decreases the available water during the spring freshet and may result in a lower freshet volume. Additionally, it increases the risk of low water supply during the warm season, threatening water supply for hydroelectricity generation (Filion 2000; Roberts *et al.* 2006), agricultural productivity (Pentney and Ohrn 2008), and contributing to or exacerbating summer drought conditions (Bonsal *et al.* 2011; Hanesiak *et al.* 2011).

This research has enhanced our knowledge of atmospheric circulation patterns conducive to snowmelt-generating above-freezing winter temperatures and rainfall in western Canada. Additionally, it has provided new insight into winter hydroclimatic conditions, particularly as it is related to persistence of atmospheric regimes through the grouping of synoptic types into similar regimes. Previous studies have evaluated trends in the frequency of synoptic types (e.g., Newton *et al.* 2014a,b; Bonsal *et al.* 2017; Bonsal and Cuell 2017); however, this study used a new approach to identify statistical step-changes in synoptic type frequency, which may be beneficial for the evaluation of thresholds related to system changes or the generation of extremes (e.g., McGregor 2017). This research has provided valuable information regarding the role of atmospheric circulation, particularly that of persistence, in winter hydroclimatic variability; however, the potential for hydrologic extremes and large-scale threats to winter snowpack merits continued research.

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Chapter 4: Hydroclimatic drivers of Mid-Winter Break-up of River Ice in Western Canada and Alaska

Abstract

The mid-winter break-up of a competent river ice cover can cause ice jamming and flooding, which can have profound impacts on the structure and strength of the ice cover. This research identifies 52 mid-winter break-up events in western Canada (1950-2008) and Alaska (1950-2014) and evaluates the hydroclimatic drivers including temperature and precipitation. The identified mid-winter break-up events are primarily located in the temperate zone, defined as the region between 400 and 1000 winter (Dec-Feb) freezing degree-days. Further delineation by terrestrial biome revealed considerable variability in hydroclimatic triggers, particularly the role of freeze-thaw days (T_{max} >0°C and T_{min} < 0°C) in Tundra and Boreal Forest/Taiga biomes and short-term (3-day) warming events in Temperate Coniferous Forests and Temperate Grasslands, Savannas, and Shrublands. The classification of 5-day sequences of mid-tropospheric circulation indicates that a persistent trough of low-pressure over Alaska and the North Pacific is the dominant pattern preceding mid-winter break-ups. Furthermore, the trough is stronger for events in British Columbia and Alberta compared with Alaska and the Yukon. The results of this research improve our understanding of the hydroclimatic conditions that generate mid-winter break-up events in western Canada and Alaska and will aid in the prediction and risk management of such events.

4.1. Introduction

The break-up of river ice is responsible for some of the most extreme flooding in cold regions (Beltaos and Prowse 2001; Prowse and Beltaos 2002; Jasek *et al.* 2003; de Rham *et al.* 2008), and can have severe impacts on geophysical, socio-economic, and biological systems (e.g. Van der Vinne *et al.* 1991; Doyle 1992; Prowse and Culp 2003; Lind *et al.* 2014). Of particular concern is the mid-winter break-up of a competent ice cover, which is often unexpected and can cause localized flooding that can be difficult to manage, especially following the return of freezing temperatures (Beltaos 2002). Furthermore, the refreezing of river ice can increase the thickness and strength of the ice cover (Beltaos 2002), and exacerbate spring break-up conditions (Beltaos 1997; Beltaos 1999; Beltaos 2008; Janowicz 2010).

The formation, growth, and break-up of river ice are important components of the hydrologic cycle in cold regions. There is a strong correlation between air temperature and the timing of river ice freeze- and break-up (Bonsal and Prowse 2003; Bieniek *et al.* 2011). Therefore, the duration and extent of the river ice season can be estimated using the timing of the 0°C isotherm, which generally increases with increasing latitude (Bennett and Prowse 2010).

River ice break-up is influenced by both thermodynamic and hydrodynamic processes (Beltaos and Prowse 2001; Beltaos 2003). In a purely thermal break-up, thermodynamic processes dominate and resisting forces, including the strength of the ice cover and attachment to riverbanks, diminish without a substantial increase in driving forces (Beltaos 1997). Thermal break-up often occurs when spring temperatures are mild and the combination of slow snowmelt and little or no rainfall produce low runoff (Gray and Prowse 1993; Beltaos 2003). Conversely, a purely mechanical break-up occurs when driving forces, caused by rising discharge and stage, increase to the point where they exceed maximum resisting forces (Beltaos 1997), often resulting in high flood stage (de Rham *et al.* 2008). Mechanical break-up is often associated with a large, rapid spring discharge pulse driven by high-intensity snowmelt, and in some cases heavy rainfall (Gray and Prowse 1993; Beltaos 2003).

The decay of an intact river ice cover is controlled by numerous energy inputs. These include: short- and long-wave radiation; latent heat of vaporization; sensible heat exchange, which is a function of the difference between air and ice surface temperatures; heat transfer from precipitation (Prowse and Marsh 1989; Hicks 2008) and warm water flowing under the ice cover providing heat to the base of the ice (Gray and Prowse 1993). Decay commences as energy inputs increase the temperature of the ice cover; once it becomes 0°C isothermal, the ice cover begins to melt (Gray and Prowse 1993; Hicks 2008). Initial melt occurs at ice crystal boundaries, causing the ice cover to fragment (Beltaos 2003). The mechanical strength of the ice cover decreases (Prowse *et al.* 1990b) and it separates from the shoreline and consequently, the ability for a spring flood wave to proceed downstream without jamming increases (Gray and Prowse 1993). Thermodynamic processes play an important role during spring as temperatures rise and the angle of solar declination increases. When the spring freshet is anomalously early, the river ice cover lacks thermal or radiative deterioration (Beltaos 2002) and the spring flood wave reaches an area with a hydraulically strong, intact ice cover that may still be attached to the bed and banks (Prowse *et al.* 1990a). These conditions can result in dynamic break-up events accompanied by ice jamming and extreme water levels (de Rham *et al.* 2008).

Winter temperatures rising above freezing and/or rain-on-snow can trigger a mid-winter break-up (Prowse *et al.* 1990b; Beltaos 2002; Beltaos and Prowse 2001; Beltaos *et al.* 2003). Mid-winter break-ups lack the thermal decay processes that occur during spring and therefore, the ice cover is typically strong and intact and break-up is mechanical (Beltaos 2002; Beltaos 2003). Unlike the spring break-up event, which clears the fragmented ice, mid-winter break-up refreezes, and the resulting ice cover is thicker and rougher, with a greater hydraulic strength (Beltaos 2002; Beltaos 2003). Notably, if the mid-winter melt is large enough to substantially reduce the volume of water equivalent in the surrounding snowpack, the potential for high hydrodynamic forces during the spring melt is decreased (Beltaos 2008).

Numerous variables influence the magnitude of the mid-winter break-up, including hydrologic parameters such as freeze-up stage, ice thickness, and discharge, and hydroclimatic factors, including air temperature and snow depth and density. For example, an equivalent warm spell later in the season can elicit a different hydrologic response compared with warming earlier in the season (Doyle and Costerton 1993). Rainfall can also be an important factor, as even modest amounts of rainfall can generate runoff. Melting snow and ice requires heat energy from the surrounding environment. Rain on snow events contribute sensible heat due to the relative warmth of the liquid water; however, the energy release through latent heat of fusion as the rain freezes in the snowpack is far greater than the energy from sensible heat (Gray and Prowse 1993).

Consequently, rainfall will generate more energy on a cold snowpack compared with an isothermal snowpack.

Mid-winter break-ups are common in temperate and maritime regions of North America, and typically occur during January and February (Beltaos 2002; Prowse *et al.* 2002; Carr and Vuyovich 2014). Evidence indicates the frequency of mid-winter break-ups is increasing in some regions, and occurring in regions that have not previously reported mid-winter events. For example, Prowse *et al.* (2002) reported an increase in mid-winter (Jan-Feb) warming events near the northern boundaries of a defined temperate region in North America, with the greatest increases occurring in western Canada and United States. Huntington *et al.* (2003) reported an increase in mid-winter break-up on the Picataquis River in Maine, primarily attributed to winter rainfall. Beltaos (2002) determined that higher proportions of precipitation falling as rain, resulting from an increased frequency of mild winter days, contributed to higher streamflow and consequently, mid-winter break-up on the Saint John River in Maine and New Brunswick. Carr and Vuyovich (2014) detected an increasing trend on the Fox and Grand Rivers, but decreasing trend on the Kankakee River, all located in the Midwest United States. Recently, Janowicz (2010) described the first mid-winter break-up event in the Yukon, occurring on the Klondike River in 2002. Additionally, research has indicated an increase in January snowmelt in southwestern Canada (Linton 2014) and an increase in the frequency and duration of winter (Jan-Mar) warm spells in western Canada (Shabbar and Bonsal 2003).

The aim of this research is to compile a database of mid-winter break-up events on rivers in western Canada and Alaska and evaluate the climatic drivers of these events.

Specifically, the dates of mid-winter break-up events are identified and corresponding daily temperature and precipitation records from the nearest climate station and/or gridded data are extracted and compared with two threshold criteria given in Prowse *et al.* (2002) and Carr and Vuyovich (2014) to determine if those thresholds are sufficient triggers of mid-winter break-up events across western Canada and Alaska, given the biogeographic and topographic variability of the study region. Additionally, sequences of dominant daily mid-tropospheric circulation patterns are identified to determine the role of atmospheric trajectory and persistence in generating mid-winter break-up events. Understanding the hydroclimatic conditions conducive to mid-winter break-up is essential for the prediction of such events and the mitigation of impacts including flooding and damage to infrastructure and property. Additionally, it will aid in determining whether the hydroclimatic triggers of mid-winter break-up have been increasing or will increase in the future.

4.2. Study Area

Western Canada – British Columbia (BC), Alberta, Saskatchewan, Manitoba, Yukon, Northwest Territories (NWT), and Nunavut – and Alaska encompass a wide range of physiographic and hydroclimatic regions. For the purposes of this research, the study region is represented by four broad terrestrial biomes (Olson *et al.* 2001; WWF 2016). Temperate Coniferous Forests cover the southern two thirds of BC and portions of western Alberta. This region is characterized by moist, mild winters, except for leeward mountain slopes, which are drier, and subalpine regions, which are cold and snowy. The southern half of the Prairie Provinces, extending from the Alberta foothills through

Manitoba is located in the Temperate Grasslands, Savannas, and Shrublands region, and is known for cold winters, low precipitation, and frequent droughts.

The Boreal Forests/Taiga region spans northern BC, Alberta, Saskatchewan, and Manitoba, southern Nunavut, most of the NWT, southern Yukon, and central Alaska. Boreal Forests/Taiga are characterized by long, cold winters and low precipitation, primarily falling as snow. Much of this region is underlain by sporadic (10-50%) to extensive (50-90%) discontinuous permafrost (Brown *et al.* 1998). Similarly, the Tundra region is known for long, cold, dry winters. This region extends from central Yukon to the southwestern coast of Alaska as well as the north Arctic coast of Alaska, Yukon, NWT, and much of Nunavut. Additionally, a narrow portion of northern BC and the Alaska panhandle is classified as Tundra, due to the steep mountainous environment that extends through southwestern Yukon and southeastern Alaska and contains some of the highest peaks in North America. Permafrost in this region ranges from sporadic discontinuous in southern Alaska/Yukon to continuous along the north coast (Brown *et al.* 1998).

Permafrost limits the hydrological connectivity within a basin as runoff is confined to the surface and seasonally thawed active layer (Woo 1986; Woo *et al.* 2008). Rainfall runoff response rates are typically high and surface ponding is common during the spring melt period (Woo 1986). Permafrost zones are generally poorly drained and wetlands are common (Olson *et al.* 2001); however, runoff response depends on soil, topography, and vegetation (Carey and Woo 2001). The winter season is characterized by negligible energy fluxes due to the high latitude environment; consequently, high temperature gradients exist between the subsurface and overlying atmosphere (Woo

1986). Cold season warming or rainfall events can affect the temperature regime of the subsurface as water refreezes to the base of the snowpack, releasing latent heat (Westermann *et al.* 2011). These ice lenses can impact runoff from subsequent rainfall or snowmelt events (Kane 1980).

Patterns of atmospheric circulation have been shown to affect winter surface climate in western Canada and Alaska. Temperature and precipitation in this region are strongly influenced by warm, moist air masses originating over the Pacific Ocean and outbreaks of cold, dry Arctic air masses. The mountainous topography of BC and Alberta is subject to orographic effects as moist air is forced to rise and precipitate, resulting in heavy snow and rainfall on the windward slopes. The leeward slopes receive less precipitation and frequently spawn Chinook winds – dry, adiabatically warmed air descending the mountains that cause a rapid, short-term increase in temperatures and, consequently, snowpack sublimation and melt (Goulding 1978). In southwestern Canada anomalously warm winter days are associated with a mid-tropospheric trough of low pressure over Alaska and the North Pacific and adjacent ridge of high pressure over western Canada (Newton *et al.* 2014). Conversely, a ridge of high pressure over the North Pacific is associated with anomalously cold, wet winter days (Newton *et al.* 2014). In Alaska, temperature and precipitation are strongly influenced by the strength and position of the Aleutian Low. For example, an Aleutian Low to the west of the Aleutian Islands is associated with anomalously warm temperatures in central, western, and northern Alaska, but anomalously cold temperatures in the Alaska panhandle (Cassano *et al.* 2011). A low-pressure system located in the Gulf of Alaska is associated with anomalously cold temperatures over most of Alaska, but anomalously warm temperatures over the panhandle (Cassano *et al.* 2011).

4.3. Data and Methods

Mid-winter break-up onset and peak water level are identified through the meticulous inspection of pen-recorder charts from 1950 to ~1996 and digital stage records from ~1996 to 2008 obtained from Water Survey of Canada (WSC) for 90 unregulated rivers in BC, Alberta, Saskatchewan, Manitoba, NWT, Yukon, and Nunavut using the criteria described in Beltaos (1990). The timing of freeze- and break-up are determined through evaluation of stage records and annotation of the WSC ‘B’ dates on daily discharge records, where ‘B’ indicates the presence of ice, but does not specify ice cover extent or thickness. Stage decreases as the river begins to freeze, and following the formation of a full ice cover the river stage stabilizes and remains relatively low. Mid-winter break-up events were identified as a ‘spike’ in stage over a short period of time.

Mid-winter break-up events in Alaska are identified using the Cold Regions Research and Engineering Laboratory (CRREL) Ice Jam Database (CRREL 2015). The ice jam database is a comprehensive record of freeze- and break-up ice jam events. The precise location of the ice jam is given, which, in many cases, is between hydrometric gauging stations, whereas the Canadian events are only detected at hydrometric stations. As this database includes only those break-up events that result in an ice jam, it may exclude minor mid-winter break-up events.

Following Prowse *et al.* (2002), a temperate region for western Canada and Alaska is defined. The temperate region represents an area where a solid, sufficiently thick (~30-50 cm) ice cover is expected to form, but that is also subject to mid-winter (Jan-Feb) warm spells. The temperate region is represented by the region bounded by 400 and 1000 accumulated mean winter (Dec-Feb) freezing degree-days (FDD) calculated using data from 210 climate stations in Canada and 660 stations in the United States. As river ice phenology is a function of temperature, accumulated FDD is commonly used to model ice growth on lakes and rivers (Michel 1971). This study employs high-resolution (0.25° × 0.25°) gridded twice-daily winter (Dec-Feb) reanalysis temperature data from the European Centre for Medium Range Weather Forecasts (ERA-Interim) for 1979-2014 (Dee *et al.* 2011) to delineate the temperate region in western Canada and Alaska. Comparisons were made between ERA-Interim and reanalysis data from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR; Kalnay *et al.* 1996) and a high-resolution (10 km × 10 km) gridded dataset using the ANUSPLIN thin-plate spline interpolation method of climate station data in Canada (McKenney *et al.* 2011). The resulting temperate region was similar between datasets; however, the NCEP/NCAR Reanalysis data was too coarse and the ANUSPLIN dataset is only available for Canada. The ERA-Interim data was deemed most appropriate for this research.

Temperature and precipitation data are obtained from the climate station nearest to the hydrometric station (Canada) or location of the ice jam (Alaska). In several cases, the nearest climate station is located outside the watershed, or could not adequately describe the hydroclimatic conditions leading up to the mid-winter break-up, and high-resolution gridded climate data from ERA-Interim is used to supplement the station data. Precipitation phase and volume are of particular concern, as point source data is not necessarily characteristic of the entire watershed, particularly for large watersheds with varying topography (Carr and Vuyovich 2014). Temperature data is used to calculate FDD prior to and following the mid-winter break-up event and any identified event not bounded by sufficient ice growth is rejected.

Although it is known that anomalously warm temperatures and/or rain-on-snow trigger the mid-winter break-up of an ice cover, few studies have quantified temperature or rainfall thresholds related to these break-up events. Two such studies, Prowse *et al.* (2002) and Carr and Vuyovich (2014) used a temperature-index approach, along with hydrometric records, to define break-up events. Specifically, Prowse *et al.* (2002) defined a minimum of 25 melting degree-days (MDD) over a 7-day period to initiate a mid-winter break-up. Carr and Vuyovich (2014) do not distinguish threshold differences between spring and mid-winter break-up events, except that an ice cover reforms following mid-winter break-up events. They found that mechanical break-up events were attributed to a MDD of 8-19 or 2.8-7.6 mm of rainfall over a 5-day period, where a thermal break-up was attributed to a MDD of 22-31 over a 5-day period. Note that the rivers evaluated by Carr and Vuyovich (2014) fall within the temperate region defined by Prowse *et al.* (2002). This research compares 7- and 5- day MDD and 5-day rainfall for each event to determine if the criteria used by Prowse *et al.* (2002) and Carr and Vuyovich (2014) is adequate to pick up the majority of events in western Canada and Alaska.

Mid-winter break-up events are often preceded by several days of anomalous winter weather. Therefore, a multi-day synoptic classification, based on a series of daily atmospheric circulation patterns (Compagnucci *et al.* 2001; Philipp 2009) rather than a daily classification, is more appropriate. The sequencing of atmospheric circulation patterns provides an enhanced understanding of the persistence and trajectory of atmospheric states and has been used to evaluate extreme temperature and hydrologic events (e.g. Jacobeit *et al.* 2006; Peña *et al.* 2015).

The method of Self-Organizing Maps (SOM) is used to classify 5-day sequences of 500 hPa geopotential heights, obtained from NCEP/NCAR (Kalnay *et al.* 1996), to identify dominant atmospheric circulation patterns preceding the mid-winter break-up events. SOM uses competitive and cooperative learning to cluster and organize data vectors (Vesanto *et al.* 2000; Kohonen 2001) and has been successful in classifying daily synoptic-scale atmospheric circulation patterns (e.g. Cassano *et al.* 2011; Newton *et al.* 2014). Sequences of 5 days were considered suitable for this research as it is long enough to indicate the spatial and temporal structure of atmospheric states, but short enough that variability is minimized. For each identified mid-winter break-up event, the corresponding daily temperature and precipitation data are examined to determine the timing of the warm spell and/or rainfall that contributed to the break-up event. In most cases, the last day of the 5-day sequence is the same as the onset of the mid-winter break-up event. However, for some mid-winter break-up events the 5-day sequence was selected to be the day prior to the break-up. In these cases, the warm spell or rainfall ended the day prior to the break-up event date, and generally occurred in larger river basins where upstream warming or rainfall triggered the break-up event. For events in Alaska, the CRREL Ice Jam Database report was reviewed to ensure adequate match between weather conditions and timing of break-up.

4.4. Results and Discussion

In western Canada, 15 of the 90 WSC hydrometric stations had at least one mid-winter break-up event between 1950 and 2008, and many of these rivers experienced multiple events. Additionally, 6 mid-winter break-up events were found in Alaska between 1950 and 2015. A total of 52 mid-winter break-up events occurred over the study period and are summarized in Table 4-1. The sizes of the drainage basins vary considerably, and there does not seem to be any discernable pattern between basin size, frequency, or timing of events.

WSC stations (Canada) and mid-winter break-up jams (Alaska) are shown in Figure 4-1, superimposed over terrestrial biomes (Olson *et al.* 2001). The four biomes in western Canada and Alaska – Temperate Coniferous Forests; Boreal Forests/Taiga; Temperate Grasslands, Savannas, and Shrublands; and Tundra – reflect the biogeography of each region, and thus hydroclimatic characteristics of each watershed. These biomes are not indicative of basin topography, which influences airflow characteristics, precipitation distribution, and runoff response. However, categorizing mid-winter break-up events by terrestrial biome is a first step in improving our understanding of the triggers of mid-winter break-up events by considering variables that affect the hydroclimate of the region (e.g. Brooks *et al.* 2013).

Table 4-1: Summary of mid-winter break-up events by Water Survey of Canada station (Canada) or location of ice jam (Alaska)

River name	Station ID*	Latitude (°)	Longitude (°)	Drainage area (km^2)	Number of events	Dates of break-up+
Canada						
Athabasca River	07AE001	54.21	-116.06	19600	2	12 Feb 1971, 21 Jan 1975
Blindman River	05CC001	52.35	-113.79	1796	2	29 Jan 1998, 10 Mar 2005
Chilcotin River	08MB005	51.85	-122.65	19200	2	14 Feb 1979, 3 Feb 2005
Coldwater River	08LG048	49.86	-120.91	316	13	24 Dec 1969, 21 Jan 1976, 26 Dec 1980, 20 Jan 1986, 11 Jan 1987, 25 Jan 1989, 28 Jan 1993, 5 Jan 1994, 31 Jan 1995, 4 Feb 1996, 28 Jan 2002, 24 Jan 2003, 24 Dec 2005
Elbow River	05BJ004	50.95	-114.57	791	1	16 Jan 1981
Fraser River	08KB001	54.01	-122.62	32400	2	5 Jan 1962, 2 Mar 1963
Klondike River	09EA003	64.04	-139.41	7810	1	15 Dec 2002
Little Smoky River	07GH002	55.46	-117.16	11100	1	26 Feb, 1992
Oldman River	05AA023	49.81	-114.18	1446	3	7 Mar 1979, 1 Jan 2000, 20 Jan 2005
Quesnel River	08KH006	52.84	-122.22	11500	4	11 Jan 1971, 11 Feb 1982, 14 Jan 2007, 30 Jan 2008
Similkameen River (Hedley)	08NL038	49.38	-120.15	5580	3	17 Jan 2001, 26 Dec 2005, 20 Jan 2007
Similkameen River (Princeton)	08NL007	49.46	-120.50	1810	3	7 Jan 1970, 23 Jan 1987, 3 Jan 2007
Smoky River	07GJ001	55.72	-117.62	50300	2	3 Dec 1956, 4 Jan 1958
Stikine River	08CF001	57.49	-131.75	36000	1	24 Feb 1992
Wapiti River	07GE001	55.07	-118.80	11300	6	26 Jan 1968, 2 Mar 1986, 29 Feb 1992, 16 Mar 1996, 12 Dec 1998, 5 Mar 2005
Alaska						
Kuskokwim River (Aniak)	15304000+	61.59	-159.55	80549+	1	13 Nov 2014
Kuskokwim River (Bethel)	15304300	60.79	-161.75	129500	1	26 Nov 2010
Naknek River	15297890	58.69	-156.66	6806	1	1 Mar 1963
Nushagak River	15302500	59.35	-157.47	25486	1	19 Nov 2003
Tanana River	15481000	64.45	-147.07	44651	1	19 Nov 2002
Wasilla Creek	15285000	61.64	-149.20	49	1	31 Dec 2010

* MWB date is the date of break-up initiation or the date of peak flow in the cases where initiation date was unavailable. For Alaska data, MWB is given as the date listed in the CRREL database
+Station ID and estimated area from Crooked Creek

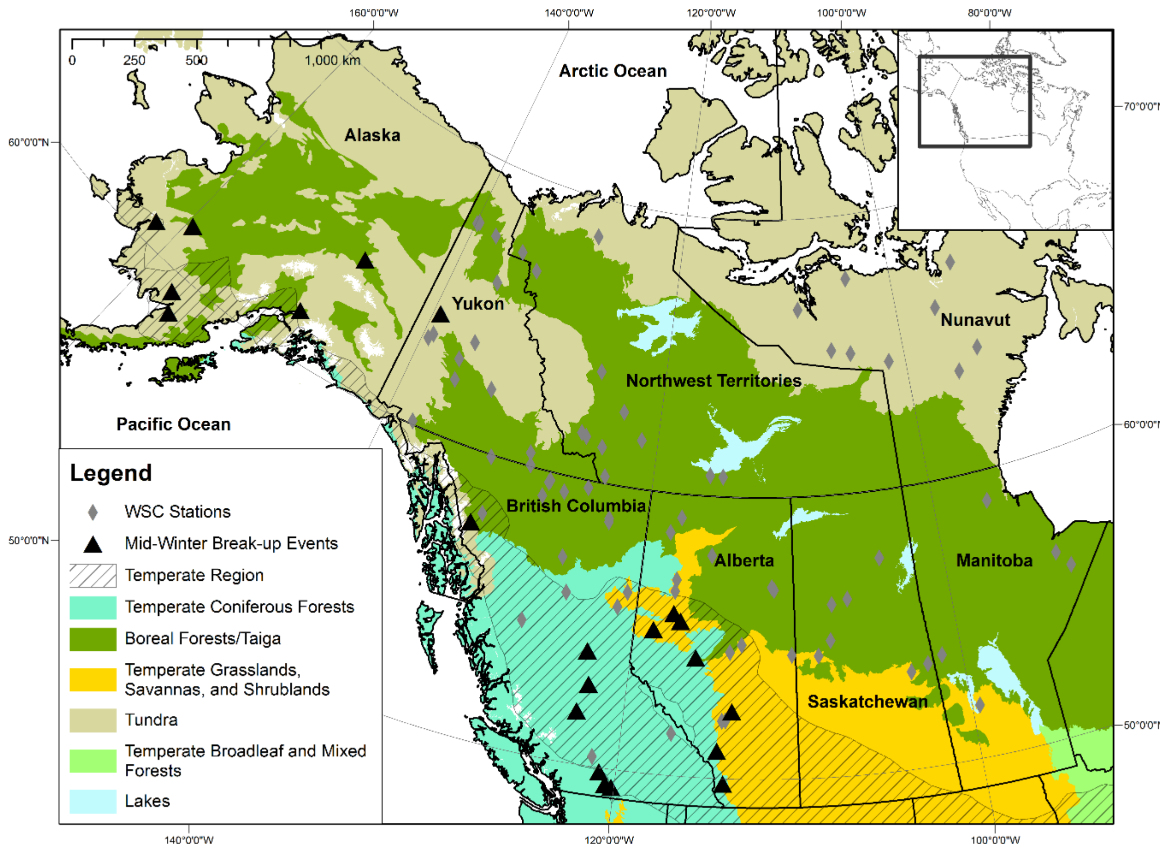

Figure 4-1: Map of the study area indicating terrestrial biome regions and showing the locations of all Water Survey of Canada (WSC) stations evaluated for this research, and locations of identified mid-winter break-up events

The temperate zone, between 400 and 1000 FDD, is consistent with Prowse *et al.* (2002); however, it extends further south in BC and is extended north through coastal Alaska. As previously noted, the NCEP/NCAR Reanalysis and ANUSPLIN datasets produced similar temperate regions. Fifteen of the 21 mid-winter break-up sites are located within the temperate region, while 2 WSC stations are located just outside of the temperate zone, the Coldwater River and Similkameen River (Princeton), and 4 locations are north of the temperate zone: the Klondike River, Tanana River, Kuskokwim River (Aniak), and Wasilla Creek. The four events associated with these northern locations are also relatively recent, all occurring within the 21st century. Additionally, these four events occurred in late autumn-early winter (Nov-Dec).

The Coldwater River has the greatest number of mid-winter break-up events with a total of 13 from 1969-2005. The Coldwater River is a tributary to the Nicola River, a basin with not only numerous mid-winter break-up events, but also several premature spring break-ups that have caused severe flooding. Doyle (1988) and Doyle (1992) describe severe break-up ice jam flooding events in January 1984 and February 1991, respectively. Both events were triggered by heavy rainfall and melting snow resulting in sequential ice jams and ice runs (Doyle 1992). Although these events occur during the winter season, the ice cover failed to reform following these break-up events; therefore, they are not classified as mid-winter break-ups for the purpose of this research.

Comparison with criteria used by Prowse *et al.* (2002) found that a minimum of 25 MDD over a 7-day period was a poor predictor of mid-winter break-up events, with only 6 of 52 events preceded by MDD of this magnitude (Table 4-2). A total of 32 events were preceded by the thresholds given by Carr and Vuyovich (2014): a minimum 8 MDD or 2.8 mm of rain over a 5-day period. The classification of mid-winter break-up events by biomes reveals the regional variability in hydroclimatic triggers. In particular, the threshold of 8 MDD or 2.8 mm of rainfall is a good predictor of mid-winter break-up events in Temperate Grasslands, but a poor predictor of mid-winter break-up events in Boreal Forests. Examination of mean and maximum temperatures preceding mid-winter break-up events revealed that above-freezing temperatures and/or rain on snow over as little as 3 days is sufficient to generate a mid-winter break-up in all hydroclimatic regions.

Table 4-2: Number of events that meet temperature and precipitation thresholds given by Prowse *et al.* (2002) and Carr and Vuyovich (2014)

	25 + MDD over 7 days (Prowse *et al.* 2002)	8+ MDD or 2.8+mm rain over 5 days (Carr and Vuyovich 2014)
All Events	6 (11%)	32 (60%)
Temperate Coniferous Forests	4 (12%)	20 (61%)
Boreal Forests	0 (0%)	1 (33%)
Temperate Grasslands	2 (18%)	8 (73%)
Tundra	0 (0%)	3 (60%)

All of the mid-winter break-up events in the Tundra region were preceded by rainfall, although some 5-day rainfall totals were relatively small (< 1 mm). However, rainfall was not a good predictive factor for events in the Boreal Forests region. The majority of the events in the Tundra, and all of the events in the Boreal Forest, were associated with a relatively high 3-day sum of daily maximum temperatures (MDD_{max}) despite low 5-day MDD. The role of maximum daily temperature compared with minimum temperature is evident in the analysis of early and late spring break-up in Alaska (Bieniek *et al.* 2011). High-latitude regions experience a large diurnal temperature range, particularly during the winter season. Therefore, examining maximum daily temperature (MDD_{max}) and freeze-thaw days ($T_{min} < 0°C$ and $T_{max} > 0°C$; Zhang *et al.* 2011) may be more appropriate.

Sequences of 5-days leading up to, and in most cases including, the day of mid-winter break-up, are classified using SOM into 3 dominant sequence patterns (Figure 4-2). Note that a larger number of sequence patterns were tested, including 6 and 4; however, these resulted in redundant sequences. Additionally, 2 sequence patterns and a sequences comprised of averages using the entire dataset were tested, but did not provide adequate variability to describe all mid-winter break-ups given the spatial distribution of events. The 3 sequence patterns are somewhat similar, which highlights the role of warm, moist Pacific air masses in triggering mid-winter break-ups, which has previously been identified as a trigger for mid-winter break-up in southern BC (Doyle and Costerton 1993). Furthermore, similarities between days within each sequence emphasize the importance of synoptic pattern persistence on extreme events.

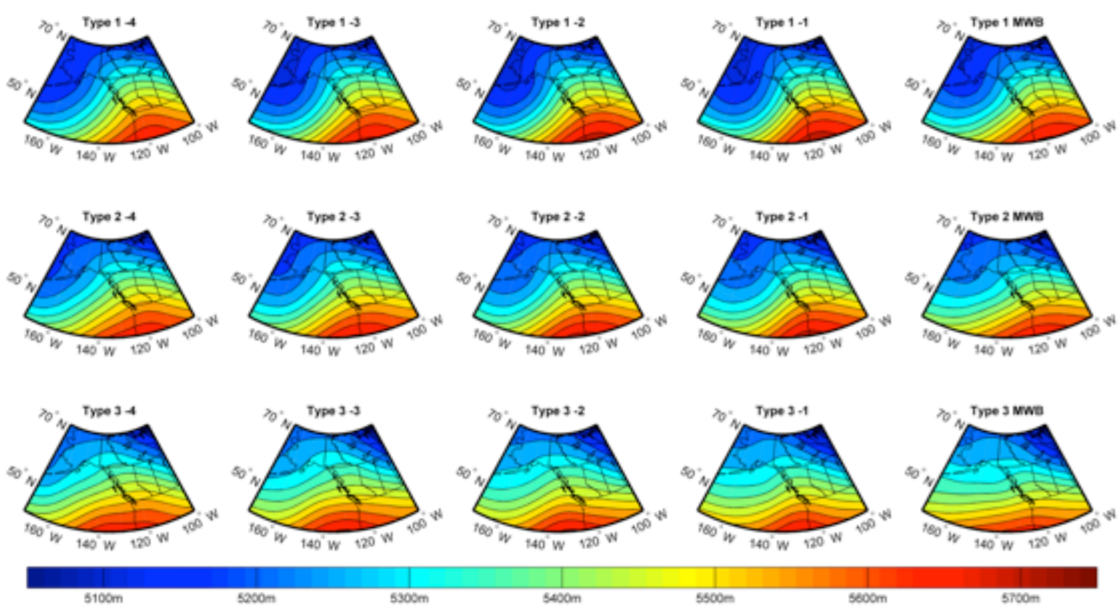

Figure 4-2: Dominant 5-day sequences of daily 500 hPa geopotential heights leading up to mid-winter break-up (MWB), classified using Self-Organizing Maps

Sequence Type 1 is characterized by a series of 5 days with a strong trough of low-pressure over Alaska and the North Pacific and an adjacent ridge of high-pressure over western Canada, which is known to be associated with anomalously high temperatures in western Canada, and has been shown to exhibit high persistence (Newton *et al.* 2014). Sequence Type 2 is similar, but with a less pronounced trough and ridge and wider spacing between contours. The trough and ridge exhibited in Sequence Type 3 is subdued, and the break-up date (MWB) of this sequence is more indicative of zonal flow. There is a noticeable expansion of higher pressure and wider spacing of contours over Alaska compared with Sequence Types 1 and 2. Meridional mid-tropospheric flow, depicted by positive phase of the Pacific North American Pattern (PNA) is associated with a strong surface Aleutian Low (Wallace and Gutzler 1981), which has been shown to influence surface temperatures in Alaska (Cassano *et al.* 2011). Specifically, anomalously high temperatures in Fairbanks and Anchorage occur when a strong Aleutian Low was positioned over the Aleutian Islands. Additionally, Bieniek *et al.* (2011) found that a 500 hPa circulation pattern similar to the structure of the positive phase of the PNA was associated with earlier spring break-up in Alaska. The frequency of each sequence is given in Table 3, subdivided into north and south. What is most striking is the dominance of Sequence Type 3 preceding those events in Alaska and the Yukon, while events in BC and Alberta were primarily associated with Sequence Type 1.

Table 4-3: Frequency of identified Sequence Types of daily 500 hPa geopotential heights given in Figure 4-2, for mid-winter break-up events in northern (Alaska-Yukon) and southern (BC-Alberta) regions

	Sequence 1	Sequence 2	Sequence 3
All	23 (44%)	12 (23%)	17 (33%)
Alaska-Yukon	1 (14%)	1 (14%)	5 (71%)
BC-Alberta	22 (49%)	11 (24%)	12 (27%)

Previous research has indicated that mid-winter break-up events can exacerbate spring break-up and ice jams (Beltaos 1997; Beltaos 1999; Beltaos 2008). Although this wasn't explicitly evaluated in this research, there is evidence that some of the mid-winter break-ups described here were associated with subsequent spring ice jams. Janowicz (2010) reported break-up flooding on the Klondike River in spring 2003 attributed to the mid-winter break-up ice jam that formed in December 2002. The CRREL database reported a severe spring break-up ice jam flood on the Tanana River at Salcha, Alaska, attributed to the ice jam that formed in November 2002 and remained in place throughout the winter (CRREL 2015). The spring ice jam flood persisted for several days and adversely affected numerous homes in the community.

4.5. Conclusions

This study identifies mid-winter break-up events in western Canada through the examination of pen-recorder and digital stage records from 90 WSC hydrometric stations from 1950-2008. A total of 15 stations were identified as having at least one mid-winter break-up, with several of those stations having multiple break-up events. Altogether, 46 events were identified across BC, Alberta, and the Yukon. Additionally, 6 mid-winter break-up events in Alaska were extracted from the CRREL Ice Jam Database. The hydroclimatic drivers of these events were examined, particularly in the context of two previous studies, Prowse *et al.* (2002) and Carr and Vuyovich (2014). The Prowse *et al.* (2002) study evaluated trends in mid-winter break-up events in a defined temperate region. Carr and Vuyovich (2014) documented a set of criteria accompanying a series of mechanical and thermal break-up events in the mid-western United States, and did not explicitly assess thresholds for mid-winter break-up, but noted that these events were defined as events followed by refreezing of the ice cover.

The majority of mid-winter break-up events fell within the temperate region, defined as the region between 400 and 1000 winter (Dec-Feb) FDD. Prowse *et al.* (2002) describe the temperate region as having both the sufficient winter ice growth and frequent warm spells necessary to generate mid-winter break-ups. The mid-winter events were further classified by terrestrial biomes, which improved the evaluation of criteria preceding the mid-winter break-up events. Although the majority of mid-winter events in the Temperate Coniferous Forests and Temperate Grasslands regions were preceded by a minimum 8 MDD or 2.8 mm of rain over a 5-day period, a 3-day warm spell and/or rain on snow was found to be a sufficient trigger. The 5-day threshold given by Carr and Vuyovich (2014) was found to be a poor predictor of mid-winter events in the Tundra and Boreal Forest regions where diurnal temperature range is high. Alternatively, the 3-day sum of maximum daily temperatures, MDD_{max}, has emerged as an important trigger in both regions, while rainfall was only a factor in the Tundra region.

Classification of 5-day sequences of mid-tropospheric circulation patterns revealed that a persistent trough of low-pressure over Alaska and the North Pacific and ridge of high-pressure over western Canada was the most common pattern leading up to mid-winter break-up events, particularly in southern BC and Alberta. Mid-winter break-up events in Alaska and the Yukon were commonly preceded by a persistent subdued mid-tropospheric trough of low pressure over the North Pacific.

This research had increased our understanding of the hydroclimatic triggers of mid-winter break-up in western Canada and Alaska through the examination of temperature, precipitation, and atmospheric patterns preceding the identified break-up events. Although classification of mid-winter events using biomes has provided insight into the regional hydroclimatic triggers of mid-winter break-up, further research that includes additional parameters such as basin size, topography, and river slope would be invaluable to our understanding of the triggers of mid-winter break-up. Future research should focus on evaluating temporal trends and future projections of 3-day warm spells (mean daily temperature above freezing), freeze-thaw days, and rain on snow. Additionally, the contribution of mid-winter break-up to extreme spring break-up should be quantitatively evaluated.

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Chapter 5: Conclusions

This dissertation presented a comprehensive analysis of the winter hydroclimatic variables and mid-tropospheric circulation patterns conducive to snowmelt in western Canada, and linked these variables to the mid-winter break-up of river ice. Snowmelt is a nonlinear response to temperature and precipitation where a threshold energy input is required to trigger a phase change from solid to liquid, which is dependent upon the antecedent conditions of the snowpack (USACE 1956; DeWalle and Rango 2008). Trends in the frequency and magnitude of winter above-freezing mean daily temperatures and rainfall from 1946-2012 were quantified and the time series at four points in key watersheds – the Fraser, Peace, Athabasca, and Saskatchewan, highlighted the interannual variability. Additionally, the average rate of change of each variable at 100 m elevation bands for each major watershed was calculated. A synoptic climatological approach was used to examine the atmospheric drivers of winter above-freezing temperatures and rainfall, with a particular focus on the role of persistence. Finally, a database of mid-winter river ice break-up events in western Canada and Alaska was created, and the hydroclimatic and atmospheric drivers of individual events were evaluated. Previous research identified winter hydroclimatic trends related to snow accumulation and melt in western Canada, including decreases in the number of frost days and increases in winter temperature (Zhang *et al.* 2000; Vincent and Mekis 2006; Zhang *et al.* 2011; Mekis and Vincent 2011; Vincent *et al.* 2015; O'Neil *et al.* 2017), and decreases in spring snow cover extent (Déry and Brown 2007; Brown and Mote 2009; Hernández-Henríquez *et al.* 2015) and snow water equivalent (Kang *et al.* 2016; Najafi *et al.* 2017). However, this is the first study to focus on temperature and precipitation trends above a threshold that is conducive to snowmelt and link these variables to large-scale mid-tropospheric circulation patterns and the generation of extreme hydrologic events.

5.1. Major Findings

The research given in Chapter 2 describes widespread significant increases in the frequency and magnitude of above-freezing temperatures in all major basins in the study area and significant increases in the frequency and magnitude of rainfall, primarily along the Pacific coast and in the upper Peace, Fraser, and Columbia river basins. Considerable spatial and temporal variability was evident across the study region. Trends were dominated by increases during March, indicating a shift toward earlier snowmelt and spring freshet. Increasing trends of lower magnitudes were also evident in January, which is particularly alarming for the potential loss of snowpack and generation of extreme hydrologic events. This research identified critical elevation zones in each major river basin where the greatest rate of increases in above-freezing temperatures and rainfall have occurred. These critical zones represent the greatest risk for winter snowmelt, loss of snowpack, and the generation of extreme hydrologic events. Previous research has primarily focused on temporal trends (e.g., Choi *et al.* 2010; O'Neil *et al.* 2017) with few studies evaluating the elevations at which the greatest changes have occurred (e.g., Brown and Mote 2009; Hernández-Henríquez *et al.* 2015). The identification of such elevation zones is an important step in understanding watershed-scale hydroclimatic processes and potential tipping points for hydrologic regime changes.

The greatest rate of change in the Fraser basin occurs below 500 m for above-freezing temperatures, and below 700 m for accumulated rainfall, which represents a small percentage of the basin, along the river valley and lower Fraser River basin near the outlet. However, increases are seen as high as 2000 m and it is evident that widespread increases have occurred throughout the basin. Similarly, the greatest rate of change in the north coastal basins occurs below 500 m for all variables except accumulated rainfall where the greatest rate of change occurs below 1000 m. The greatest rate of change in the frequency of above-freezing temperatures and accumulated MDD in the Peace, Athabasca, Saskatchewan, and Columbia river basins occurs between 1000 and 1500 m. For the Peace, Athabasca, and Saskatchewan river basins this elevation represents the eastern foothills and lower-elevation areas of the Rocky Mountain headwaters, a region that supplies a large portion of the annual streamflow on these rivers. Critical elevation zones with respect to rainfall are more variable, but for the Peace, Athabasca, and Saskatchewan river basins, these zones are all concentrated in the foothills and low- to mid-elevation Rocky Mountain headwaters.

The rates of changes in all river basins are related to the temperature climatology at each elevation band. In the north coastal, Fraser, and Columbia river basins the warmest region is at low elevations. In the Peace, Athabasca, and Saskatchewan river basins the region between 1000 and 1500 m has the highest mean winter temperature, which is the foothills region east of the Rocky Mountains. Under continued climate change additional elevation bands may become more vulnerable to periods of above-freezing temperatures and rainfall. In the north coastal, Fraser, and Columbia river basins the temperature decreases with increasing elevation; therefore, vulnerability will likely parallel elevation. Conversely, the mean winter temperature in the Peace, Athabasca, and Saskatchewan river basins steeply rises with increasing elevation and decreases steadily above 1500 m; therefore, higher elevation regions will likely become more vulnerable. Mid-winter thawing in steeper environments will likely have a faster runoff response and increased potential to trigger downstream flooding and ice break-up. Given the importance of high elevation mountain snowpack to water resource availability in these watersheds, increased vulnerability will likely threaten the reliability of water resources and require enhanced water resource management strategies.

Chapter 3 presented the synoptic-scale mid-tropospheric drivers of winter above-freezing temperatures and associated rainfall. The classification of daily mid-tropospheric geopotential heights (GPH) was achieved using Self-Organizing Maps (SOM). SOM is particularly advantageous for synoptic typing as the topological organization of the SOM array facilitates the analysis of particular features of atmospheric circulation such as trajectory and identification of the evolution of atmospheric states. The most dominant – frequent and persistent – patterns occur in the four corners of the SOM array with transitory patterns in between. The topological ordering of the SOM array also facilitates the grouping of patterns into synoptic type regimes, which proved useful for examining the role of persistence in winter surface climate response. Results reveal that the frequency of days when the mean daily temperature is above freezing and days with rainfall are both higher and more spatially widespread during days when the mid-troposphere is characterized by a ridge of high pressure over western Canada compared with zonal flow or a ridge of high pressure over the Pacific Ocean and adjacent trough of low pressure over western Canada. Zonal flow is typically associated with higher precipitation; however, zonal flow is associated with a low frequency of above-freezing temperatures, therefore, precipitation falls as snow rather than rain. Under future climate warming, there may be a shift toward an increased frequency and magnitude of rainfall in western Canada during days characterized by zonal mid-tropospheric flow.

The winter climate in much of the north coastal, Fraser, and Columbia river basins is influenced by moist air masses originating over the Pacific Ocean and, consequently, the snowpack tends to be warm and wet (Mock and Birkeland 2000). Conversely, the Rocky Mountain headwaters of the Liard, Peace, Athabasca, and Saskatchewan river basins are influenced by continental and Arctic air masses and tend to have cold, dry snowpacks (Mock and Birkeland 2000). A wet snowpack generally produces a faster runoff compared to a dry snowpack (Colbeck 1975; Singh *et al.* 1997); therefore, the hydrologic response to a period of above-freezing temperatures or rainfall is expected to be quicker in the north coastal, Fraser, and Columbia river basins compared to the Liard, Peace, Athabasca, and Saskatchewan river basins. The Liard, Peace, Athabasca, and Saskatchewan river basins may require a greater energy input to trigger sufficient snowmelt to generate runoff, depending on the antecedent conditions of the snowpack. Energy inputs through above-freezing temperatures or rainfall may contribute to snow metamorphism, altering the snowpack (Colbeck 1975; Singh *et al.* 1997), increasing the risk of snow avalanches (Fitzharris 1987; Hägeli and McClung 2003).

Trend analysis revealed significant decreases in both zonal flow (Type 3) and a mid-tropospheric ridge of high pressure over the Pacific Ocean and adjacent trough of low pressure over western Canada (Type 1). Significant increases in a strong ridge of high pressure over western Canada (Type 10) were also detected. This suggests a long-term regime shift that is conducive to fewer colder days and more days when the mean daily temperature is above freezing and rainfall occurs, particularly in the north coastal, upper Peace, Fraser, and Columbia river basins. These trends are consistent with the significant increasing trends in the frequency and magnitude of above-freezing temperatures and rainfall described in Chapter 2. Persistent meridional flow is associated with extreme weather phenomena (Francis and Vavrus 2012; Petoukhov *et al.* 2013; Screen and Simmonds 2014), and an increase in the frequency and persistence of a ridge of high pressure over western Canada (Type 10) signifies an increased risk of sufficient energy input to the snowpack to generate snowmelt and trigger an extreme hydrologic event.

Change point analysis indicated a statistical shift in the mean, standard deviation, and slope of the distribution of the dominant zonal flow pattern (decrease; Type 3) and a ridge of high pressure over western Canada (increase; Type 10), and a step increase in the mean persistence of Type 10 in 1977, coinciding with a shift to a positive phase of the PDO (Mantua *et al.* 1997). The presence of step changes suggest a nonlinear, abrupt shift in dominant mid-tropospheric circulation patterns rather than a linear change. It is likely that step-changes in hydroclimatic and hydrologic systems parallel the changes seen in atmospheric circulation; however, further analysis is required to determine the presence and extent of these system responses.

There is potential for an immediate, high-impact consequence of a period of above-freezing temperatures and associated rainfall. One such consequence, the mid-winter break-up of river ice, was evaluated and presented in Chapter 4. A database of 52 mid-winter river ice break-up events was created including 46 events identified in western Canada from 1950-2008 and six events in Alaska from 1950-2014. These events were classified by biogeoclimatic zones and the hydroclimatic drivers were evaluated to determine the general hydroclimatic triggers for each zone and to compare triggers among zones. Results revealed spatial diversity of hydroclimatic drivers among the triggers of mid-winter river ice break-up events. Events in the Temperate Coniferous Forests (primarily in southern BC) were associated with above-freezing temperatures and rain-on-snow while events in the Temperate Grasslands, Savannas, and Shrublands zone (east of the Rocky Mountains) were commonly triggered by above-freezing temperatures. In northern regions the diurnal temperature range is quite high, and events in this area were linked to above-freezing maximum daily temperatures (Boreal Forest) or small amounts of rain-on-snow (Tundra). Additionally, for all events, three days of above-freezing temperatures or rain-on-snow was found to be a sufficient trigger, suggesting a relatively quick runoff response.

Due to the infrequent, high-magnitude nature of extreme hydrologic events it is difficult to statistically evaluate trends to determine if these events have increased linearly or nonlinearly in frequency or magnitude over time. However, it is possible to statistically evaluate the trends in the hydroclimatic drivers – above-freezing temperatures and rainfall, as demonstrated in Chapter 2. Of the 52 mid-winter river ice break-up events identified in western Canada and Alaska, 25 occurred in January, consistent with the statistically significant increases in above-freezing temperatures and rainfall during January that were identified in Chapter 2. Comparatively, four events occurred in November, all of which were in Alaska, eight events occurred in each of December and February, and seven events in March. Continued climate warming could increase the frequency of mid-winter river ice break-up events and trigger events further north and at higher elevations.

The atmospheric drivers of the mid-winter river ice break-up events were evaluated through the classification of 5-day sequences of mid-tropospheric circulation patterns leading up to each event using SOM. A persistent ridge of high pressure over western Canada, of varying strengths and positions of the ridge axis, dominated the days leading up to the mid-winter river ice break-up events. This is consistent with research presented in Chapter 3 that found a ridge of high pressure over western Canada was highly persistent, frequent, and associated with a high frequency of above-freezing temperatures and rainfall compared with other circulation types. These results highlight the role of persistence in the generation of snowmelt and extreme hydrologic events and suggest that continued increases in the frequency and persistence of a ridge of high pressure over western Canada could increase the frequency and magnitude of mid-winter river ice break-up events.

5.2. Future Research Directions

This dissertation addressed key research gaps and provided a novel perspective on winter hydroclimatic trends and variability in western Canada. Snow accumulation and melt are important components of the hydrologic cycle in cold regions, supplying the majority of the annual flow on many major rivers in western Canada. The loss of winter snowpack can have profound consequences for the generation of hydroelectricity (Cohen *et al.* 2000; Filion 2000; Peters and Prowse 2001; Martz *et al.* 2007), the oil and gas and mining industries (Prowse *et al.* 2009; Pearce *et al.* 2011; Instanes *et al.* 2016), agricultural productivity (Martz *et al.* 2007; Pentney and Ohrn 2008; Kerkhoven and Gan 2011), and aquatic ecosystems (Wrona *et al.* 2006; Burn *et al.* 2008; Wrona *et al.* 2016), and increases complexity for the management of water resources. As such, it is critical to continue hydroclimatic research and monitoring in this region.

This research has been limited by the lack of comparison with hydrometric and snow survey data. Spatial and temporal trends in hydroclimatic variables conducive to snowmelt in western Canada were evaluated and critical elevation zones in major watersheds where the greatest rate of change has occurred were identified. Comparing these results with trends and variability of snow water equivalent (SWE) using historic snow survey data in the mountainous regions of the watersheds and streamflow records will provide a critical link between snowmelt-generating hydroclimatic variables and hydrologic response. Of particular concern are the headwaters of the Peace, Athabasca, and Saskatchewan rivers. The majority of annual streamflow on these rivers originates as snowpack in mountain headwaters, a region identified as having the greatest rate of change in hydroclimatic conditions conducive to snowmelt. Scaling down research to the watershed level and focusing on case studies will improve our understanding of the relationships between atmospheric circulation, hydroclimatic trends, and surface impacts.

This research provided an innovative approach to evaluating synoptic-scale mid-tropospheric circulation patterns, particularly step-changes in the frequency and persistence of individual synoptic types and synoptic regimes. In particular, this research identified a step-change increase in both the frequency and persistence of a ridge of high pressure with a ridge axis centred near the coast, signifying an increased potential for snowmelt and the generation of extreme hydrologic events. Furthermore, this research linked the mid-winter break-up of river ice to a period of above-freezing temperatures and in some cases rain-on-snow, and a persistent ridge of high pressure over western Canada. There is a growing concern over changes to the frequency and intensity of extreme weather phenomena related to increasing Rossby wave amplitude, which favours slower moving, persistent circulation patterns (Francis and Vavrus 2012; Screen and Simmonds 2013; Peings and Magnusdottir 2014). Efforts should be made to improve our understanding of the surface climate and hydrologic impacts of persistent or amplified atmospheric circulation patterns and determine if there are thresholds related to these characteristics that trigger hydroclimatic responses. Both step-changes and changes to persistence of atmospheric or hydroclimatic variables may be key to evaluating regime shifts or tipping points in nonlinear hydroclimatic and hydrologic systems. Additionally, given the importance of persistence to the generation of extreme hydrologic events, it is recommended that future research focus on improving methods for evaluating the persistence of atmospheric circulation patterns.

Arguably, the greatest advances in hydroclimatic research have occurred through the synthesis of comprehensive research projects (e.g., Prowse *et al.* 2015a,b). As such, hydroclimatic research should be incorporated into a comprehensive, integrated watershed research program. Recently, the concept of changing hydroclimatic regimes producing 'water-rich' and 'water-poor' regions has gained interest (e.g. IPCC 2013; Prowse *et al.* 2013). Results of this research can provide meaningful input to such analyses of changing freshwater resource distribution. This research has synthesized three important research phases to evaluate winter hydroclimatic variables conducive to snowmelt and the generation of extreme hydrologic events; however, there is great potential to incorporate additional analyses to gain a more comprehensive understanding of changing winter hydroclimatic regimes.

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