Recent changes in patterns of western Canadian river flow and association with climatic drivers: A CROCWR component

by

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Abstract

Climatic variability and change can have profound impacts on the hydrologic regime of a watershed, especially in regions that are particularly sensitive to changes in climate, such as the northern latitudes and alpine-fed regions of western Canada. Quantifying historical spatial and temporal changes in hydrological data can provide useful information as to how water resources are affected by climate, as well as create an understanding of potential future variability in the hydrologic regime of a region. The CROCWR (Climatic Redistribution of Canadian Water Resources) project was established to quantify changes in western Canadian water resources under past, present, and future climate through spatio-temporal analyses of runoff and its driving climatic and atmospheric forcings.

This research involved the examination of trends in western Canadian annual and seasonal streamflow volume and timing for the periods of 1976-2010 and 1966-2010. Runoff was found to have increased significantly in the most northern watersheds studied, while mid-latitude water availability has decreased considerably. In addition, the onset of the spring freshet has shifted toward earlier timing in the North and along the Pacific coast, associated with increased freshet length and flow volume, while contrasting later freshets have occurred in the mid-latitudes, causing decreased warm season river flows in this region. Application of a Principal Component Analysis revealed coherent hydrological variability in each of the northern, mid-latitude, and southern regions of the study area, with consistent increasing and decreasing trends in river flows for the north and mid-latitudes, respectively. The results of this analysis suggest a northward shift in water from adjacent more southerly western Canadian watersheds.

Lower- and mid-latitude runoff was shown to be positively correlated with precipitation both annually and during the warm season, while the effect of temperature was found to be associated with the timing of the spring freshet in the North and along the west coast. River flows in some watersheds were shown to be influenced by the effects of the Pacific Decadal Oscillation and/or the Pacific North American low-frequency climate patterns, however, the overall influence of these natural oscillations on western Canadian streamflow was not determined to be indicative of overall trend results.

The results of this analysis will provide water resource managers with an indication of the direction and magnitude of changing water availability in western and northern Canada.

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Dedication

To Stew, Mom and Dad

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Chapter 1

Introduction

1.1 Climate, water and human society

Hydroclimatology is the study of how climate causes time and space variations (both global and local) in the hydrologic cycle. Changes in the relationship between climate and hydrology can trigger extreme hydroclimatic events, such as floods and droughts, as well as underlie future influences of global warming on water resources (Shelton, 2009). Research concerning the links between climate and hydrology has garnered much attention over the past several decades owing to the accelerated rate of anthropogenic global warming that has occurred over the past 50 to 100 years (IPCC, 2007a).

Climate change can have critical impacts on the hydrological regime of a watershed, including both production of and access to freshwater resources. The hydrologic cycle defines the potential energy of a watershed as the amount of runoff generated within the basin. The water balance equation links streamflow (R), precipitation (P), evaporation (E) (or evapotranspitation, ET), losses to regional groundwater (G), and changes in watershed storage (ΔS) through the following equation:

$$R = P - E \pm \Delta G \pm \Delta S \tag{1.1}$$

where variables are measured in water depth equivalent (mm) over time. Groundwater is water that enters and leaves the watershed subsurface, and thus does not contribute to streamflow. Water storage bodies include snow, surface water (e.g., lakes, ponds, or depressions), glacial ice, soil moisture, and groundwater (Winkler *et al.*, 2010).

Given that precipitation and evaporation are major components driving a watershed's hydrologic production, it is no surprise that global climate change is expected to alter the hydrologic cycle, placing stress on water supplies and demands (Gleick *et al.*, 2010). Evaporation is largely dependent on air temperature, where warmer temperatures generally tend to result in higher rates of evaporation. Surface air temperatures have been increasing over at least the period of record (IPCC, 2007a), and it is now well established that human activities are contributing to modern climate change by altering the natural composition of the atmosphere (IPCC, 2007a; Min *et al.*, 2008). Changes in surface temperatures have also been related to changes in atmospheric circulation and precipitation patterns (Booth *et al.*, 2012). Whereas precipitation may be dominated by snowfall in a colder climate, warming temperatures may result in a shift in the type of precipitation from snow to rain, resulting in decreased snowpacks (and hence less watershed storage in the form of snowpack) and reduced flows later in the year (in the absence of snowmelt). There is a finite supply of water that moves within the hydrologic cycle of a watershed, hence understanding the role of climate is critical to ensuring a sustainable supply of water within the water cycle (Conservation Ontario, 2010).

1.1.1 Western Canadian water resources

In addition to the effects of temperature warming and precipitation changes, water resources are coming under increasing pressure from societal demands related in large part to growing industries and rising populations. One region where this is of particular concern is western Canada. Agriculture in southern British Columbia (BC), Alberta, and Saskatchewan, shale gas production in northern BC, and oil and gas developments in northern Alberta all rely heavily on large quantities of water to employ their operations. The petroleum industry, for example, accounts for 92% of licensed surface water use in central Alberta's Athabasca River Basin, approximately 35.5% of which was actually used in 2005. Under current industrial growth, however, the sector's use of surface water has been forecasted to increase by between 120% and 165% of its 2005 usage by 2025 (Mannix *et al.*, 2010). Increases in surface water use could be of particular importance during the winter low flow months given the small capacity for storage in the Athabasca River Basin upstream of Lake Athabasca (Bawden *et al.*, 2013). Urban issues in growing cities, such as increased municipal demand, storm water mitigation, and infrastructure deficits, place further stresses on water resources in western Canada (Sommerfeld, 2012).

With a total of 6.5% of the global renewable supply of freshwater and only 0.5% of the global population, Canada is often considered to be a "water-rich" country. Unfortunately, the majority of our freshwater is not situated in places where it is needed. For example, while there is an abundance of freshwater in northern Canada, an estimated 75 percent of Canadians live much further south – within 161 kilometers (100 miles) of the U.S. border (National Geographic, 2013). In addition, water that is located within closer proximity to civilization is often not readily accessible due to it being trapped in a frozen state (i.e., as snow or in glaciers). Water supplies are spread out in different ways across the country and water challenges vary greatly.

Even across western Canada, water is not uniform in quantity. Although BC's Pacific coast rainforests and snow-covered mountains are not likely to experience water scarcities any time soon, water shortage is something policymakers and water resources managers must deal with on a daily basis just 400 km inland in the Okanagan region. Lake-dense Manitoba is prone to destructive flooding (e.g., the Red River flood of 1997), while southern Alberta is highly susceptible to drought. The north-flowing Athabasca, Peace, Liard, Slave, Mackenzie, Churchill, and Nelson rivers make the northern portions of BC, Alberta, Saskatchewan, and Manitoba, as well as the Yukon and Northwest Territories land masses, rich in water resources, yet in the south, where the majority of western Canadians live, water is not as plentiful (Sommerfeld, 2012).

The most notable supply of freshwater in western Canada is derived from glacial melt and annual runoff from the Rocky Mountains. The prairie provinces of Alberta and Saskatchewan (and eventually Manitoba) are supplied with freshwater via 1,300 glaciers on the eastern side of the Rocky Mountains, while glaciers on the western slopes provide water to BC. Eleven million people, or nearly one third of Canada's total population, rely on rivers flowing from, and/or lakes sourced by, Rocky Mountain glaciers, while almost no water originates in eastern Alberta, Saskatchewan, or western Manitoba (Sommerfeld, 2012).

One of the largest concerns regarding western Canada's current and future water supply is the rapid rate of glacial retreat since the Little Ice Age of the 19^{th} century, and more recently since the 1980s (Moore *et al.*, 2009). A combination of climate change linked to greenhouse gas (GHG) emissions and long-term climatic and atmospheric cycles are thought to be the reasons behind western Canadian glaciers now approaching their lowest levels in 10,000 years (Hipel *et al.*, 2011), and the impacts of such shrinkage could be severe. For example, the Bow Glacier, which feeds the Bow River, the main water supply to the city of Calgary, has shrunk by 27% in the last 60 years and there is speculation it could vanish entirely within the next 40 years (Hipel *et al.*, 2011). If this were to happen, Calgary – a city of over one million residents – would have significant trouble providing water to its citizens.

On top of glacial concerns, western Canadian snowpack is diminishing, reducing the amount of water stored over the winter months and consequently the amount of runoff produced in the spring and summer. Both glacial retreat and shrinking snowpacks have been accompanied by changes in runoff patterns and streamflow timing, two factors that can have substantial effects on aquatic ecosystems and urban water systems (Sommerfeld, 2012). Statistically detectable declines in late-summer streamflow from glacier-fed catchments have been observed over much of the western Canadian mid-latitudes, while glacier-fed streams in northwest BC and southwest Yukon have experienced increasing flows (Moore *et al.*, 2009). Given the lack of uniformity that already exists in western Canada's "waterscape", these changes are of utmost concern to water resources managers. Streamflow in the Mackenzie River, for example, has been predicted to increase by 20% (of its 1980-1999 level) by 2099 due to climate change (Bates *et al.*, 2008). Current mean annual flow in the Mackenzie River (at Norman Wells) is 8,500 m^3/s (Environment Canada, 2013a), hence an increase of 20% would mean an extra 1,700 m^3/s in the Mackenzie – or almost seven times the mean annual streamflow of the South Saskatchewan River (at Saskatoon) in the absence of projected drying! Dams have helped address some of the challenges resulting from non-uniform freshwater distribution and changing water patterns through diversion and/or storage of water when it is plentiful and release when it is needed, however storing water for future use without addressing supply concerns is not a sustainable long-term solution (Sommerfeld, 2012).

Whatever the cause, climate change and the related shrinking of western Canadian glaciers and snowpack will have a direct impact on future water supply. As population and economic activity continue to grow in the West, water scarcity will likely increase, at least in the places where it is needed most.

1.2 Research objectives and the CROCWR project

Though a number of hydroclimatological studies with a focus on western Canada have shown recent climate-driven changes in hydrology (see *Section 2.1.2.1*), most studies have been isolated to one or a few western Canadian watersheds or sub-regions, such as “the North” (e.g., Fleming and Clarke, 2003; Déry and Wood, 2005) or “the Prairies” (e.g., Schindler and Donahue, 2006; Bonsal *et al.*, 2013). Knowledge of how water resources and availability in western Canada as a conterminous region have changed is still lacking. Given projected changes in temperature, precipitation, and flow patterns, stronger contrasts between “water-rich” and “water-poor” regions are expected that may ultimately lead to, for example, a demand for reconsideration of inter-regional transfers of water – something that could be considered a form of climate change adaptation. There is therefore a need to conduct more detailed assessments of water redistribution over the entirety of western Canada, an area of highly contrasting hydroclimatic regimes and overlapping water-use and jurisdictional boundaries.

The Climatic Redistribution of western Canadian Water Resources (CROCWR) project involves the development of methods and tools for quantifying water resources in western Canada under past, current, and future climate regimes by taking into account the major climatic factors that have historically and will in the near future control the source, supply, and spatial redirection of intercepted water and subsequent discharge in this regime. The project includes analyses of changes and trends in runoff patterns and timing of significant hydrological events (see also Ahmed *et al.*, 2013); near-surface climatic drivers (including temperature and precipitation) and snow accumulation and melt (Linton, 2013); and large-scale synoptic and atmospheric patterns (Newton, 2013).

This portion of the CROCWR project is concerned with quantifying historical spatial and temporal trends and patterns in western Canadian streamflow, and investigating the impact that climate has had on streamflow. Specifially, the research aims to:

1. a) Assess trends in runoff in a collection of western Canadian watersheds for the four “traditional” seasons, six-month cold and warm seasons, and annually;

   b) Assess trends in streamflow timing variables at a collection of western Canadian streamflow gauges;
2. Identify regions of coherent streamflow behaviour through retrieval of the principal modes of variability in runoff and streamflow timing;
3. a) Examine linkages between regional climate and runoff/streamflow timing variables;

   b) Examine linkages (teleconnections) between large-scale atmospheric patterns and runoff/streamflow variables.

The development of the hydroclimatic modelling and the results of this analysis are intended to provide government and industry water resources managers and policy makers with new tools essential to the management of western Canadian watersheds.

1.2.1 Thesis organization

The contents of this thesis are organized are follows:

- *Chapter 2* provides a focused review of the current state of knowledge of global climate change and variability, and the links between climate and hydrology in western Canada. An overview of some of the relevant data sources and research methods used in hydroclimatology is presented.
- *Chapter 3* provides a detailed overview of western Canada, including the CROCWR study region. Climatic regions and major western Canadian drainage basins are introduced.
- *Chapter 4* details the data sources and methodologies employed in this work;
- *Chapter 5* presents the results of the study and discusses outcomes in a broader context;
- *Chapter 6* critically evaluates the results, draws general conclusions, assesses the limitations of the study, and makes recommendations for future research.

Chapter 2

Literature Review

This chapter presents an overview of global climate change and variability over the past century, including a review of recently documented trends in temperature, precipitation, and snow cover, and how these variables act as driving forces in the redistribution of water resources. The focus of this review is directed at western Canada, an important region in which the meteorological and hydrological responses of climate change can be assessed at various latitudes. Results of Canadian research have revealed high sensitivity to changes in climate in the western and northern parts of the country, hence this region is the focus of this research.

Section 2.1 discusses recent trends in both global and Canadian temperature and precipitation, Canadian snow cover, and Canadian hydrology. *Section 2.2* presents an overview of the data sets and methods commonly used in hydroclimatic research. This includes a review of the various types of data used to quantify water resources and a description of some of the different approaches used for trend detection, regionalization, and statistical linkage of hydrology and climate. A summary of the literature and a foreword to the next chapters are given in *Section 2.3*.

2.1 Recent Changes in Climate and Hydrology

2.1.1 Temperature and Precipitation

Climate change and global warming have been topics of intense research and discussion over the past several decades. Recent analyses indicate that the global mean surface air temperature rose an estimated 0.74°C since the early 20th century (0.56 to 0.92°C), and that the 50-year linear warming trend from 1956 to 2005 was nearly twice that for the 100 years from 1906 to 2005 (IPCC, 2007a). In addition, new paleo-climate data indicate that both surface air temperatures and rate of warming were greatest during the 20th century relative to any other era within the past 1,000 years (Serreze *et al.*, 2000; Moritz *et al.*, 2002; IPCC, 2007a). Driven by a desire to both understand and mitigate the impacts of climate change on human society and the environment, a number of recent of studies have contributed to our understanding of the global climate system through analyses of climate data, including various measures of temperature and precipitation, for different regions of the world.

Jones *et al.* (1999) related increasing global temperatures over the 20th century with shrinking areas of exceptionally cold climate. In particular, warming was shown to have been greatest over the normally cold high latitudes of the northern continents, predominantly during the winter and spring months. Serreze *et al.* (2000) likewise found positive trends in surface air temperatures over northern Eurasia and North America, with partly-compensating negative trends over eastern Canada, southern Greenland, and the northern North Atlantic. Temperature increases were again shown to have been strongest during the winter and spring. The Intergovernmental Panel on Climate Change (IPCC) (IPCC, 2007a) reported that Arctic air temperatures have increased at a rate of almost twice the global average since 1906.

High-latitude temperature trends have been accompanied by increased cyclonic activity north of 60°N for all seasons except autumn, and increased cyclone intensity during all seasons (Serreze *et al.*, 2000). Pronounced increases in precipitation have also been detected for the 55-85°N latitudinal band, particularly during the fall and winter, since the second half of the 20th century (Serreze *et al.*, 2000).

The pronounced recent rise in high latitude Northern Hemisphere temperature and precipitation reflects shifts in both atmospheric and anthropogenic forcings (Serreze *et al.*, 2000). Moritz *et al.* (2002), for example, determined that the Arctic Oscillation (AO) mode of atmospheric circulation played a key role in climatic changes in both the Canadian and Eurasian Arctic regions during the last 30 years of the 20th century. Mantua and Hare (2002) determined that the 1976/1977 regime shift of the Pacific Decadal Oscillation (PDO) was largely responsible for increased temperatures over the North Pacific, and specifically over northwestern North America, since this time. Zhang *et al.* (2007) and Min *et al.* (2008), on the other hand, examined the influence of anthropogenic factors on Arctic precipitation patterns and demonstrated that elevated greenhouse gas (GHG) emissions (as well as sulfate aerosols, in the latter case) have contributed to high-latitude (north of 50°N) precipitation increases, and related drying in the Northern Hemisphere subtropics and tropics, that cannot be explained by natural climate variability.

A number of studies have applied climate models to project changes in future climate worldwide. Results overwhelmingly indicate a continuation of the trend toward more intense Arctic climate. In particular, the IPCC 2007 report on climate change (IPCC, 2007a) projected a warming of about 0.2°C per decade over the next twenty years for a range of GHG emission scenarios. Warming is expected to be greatest over land and at most high northern latitudes, and least over the Southern Ocean (ocean waters south of 60°S latitude and encircling Antarctica) and parts of the North Atlantic Ocean. Yin (2005) projected a consistent poleward and upward shift and intensification of the mid-latitude storm tracks, accompanied by poleward shifts in precipitation and other climatic forces, through an analysis of several 21st century climate simulations. Salathé (2006) detected a similar northward shift and intensification of winter precipitation in 21st century climate models focused on the Pacific coast of the United States and southern Canada. Bates *et al.* (2008) indicated that precipitation increases of at least 20% are to be expected in the high latitudes of the northern and southern hemispheres, whereas severe drying is forecasted for the mid-latitudes. For Arctic watersheds, Min *et al.* (2008) found that model-simulated precipitation responses to human forcings are weaker than what actual observations have shown, implying that model projections of future precipitation may also be underestimated; this would have important implications for the development of climate change adaptation strategies.

Results tend to agree that the anthropogenic climate change signal is strongest in the northern high-latitudes, where polar amplification due to meridional heat transport and positive ice-albedo feedback results in more rapid warming than at lower latitudes (Min *et al.*, 2008). Detecting regional effects of climate change may therefore be facilitated by studying countries that span both high and mid-latitudes, such as Canada (Zhang *et al.*, 2000).

2.1.1.1 Climate Change in Canada

Zhang *et al.* (2000) studied 20th century Canadian climate records and found a distinct pattern of temperature change, including significant warming over the south and west and cooling in the northeast, over the latter half of the century. For the century-long record, mean annual temperature south of 60°N was found to have risen between 0.5 to 1.5°C, and warming of minimum temperatures was greater than that of maximum temperatures. Precipitation was determined to have increased between 5 to 35% over the second half of the century, with greatest intensification over the high latitudes; the ratio of snowfall to total precipitation was also shown to have increased due to greater total winter precipitation. Whitfield and Cannon (2000) compared meteorological data from two different 20th century decades (1976-1985 and 1986-1995) and found varying climatic responses in different regions of Canada for the two decades. Temperatures were shown to have increased in the more recent decade, primarily in the west and north, whereas winters were generally cooler in the Maritimes and southern territories. Fall and winter precipitation increased over western and northern Canada, whereas eastern Canada showed decreases and the Prairies showed mixed results. Stone *et al.* (1999) examined trends in the frequency and intensity of Canadian precipitation, and found a shift to more extreme precipitation events, particularly in the northern areas of the country. Déry and Wood (2005), on the other hand, found a consistent decline in observed precipitation over northern Canada between 1964 and 2000. Clearly, the direction and magnitude of surface temperature changes are much more consistent among climate models than are precipitation changes (Barnett *et al.*, 2005).

For western North America, Booth *et al.* (2012) studied climatic changes from 1950-2005 and found a general warming trend over the entire study region, with greatest increases along the North American Cordillera. Moderate increases in precipitation volume and intensity were also noted. Nelson *et al.* (2010) examined a 6,000-year record of changing water balance in the Pacific Northwest and discovered that, while low-frequency drought and pluvial cycles are a persistent feature of regional climate, the average duration and corresponding impact of these multi-decadal wet/dry cycles has increased over the last 1,000 years. Intensifications of the El Nino-Southern Oscillation (ENSO) and PDO over this period are suggested to have driven changes in the tropical and extra-tropical Pacific.

2.1.1.2 Snowpack and Snow Cover

Climate variability and change can impact the local environment in many ways. In cooler regions, snow depth and snow cover are highly influenced by changes in both temperature and type and amount of precipitation, thus snow cover is considered to be a useful indicator of climate change (Brown and Braaten, 1997). Snow cover is often measured as snow-water equivalent (SWE), which refers to the amount of water stored in a snowpack that would be available upon melting (Farmer *et al.*, 2009).

Brown and Braaten (1997) studied variability of Canadian monthly snow depth and snow cover from 1946 to 1995. Results indicated that the majority of Canada experienced significant decreases in winter and early spring snow pack, especially in February and March, with an abrupt transition to lower snow depth coinciding with the well-documented mid-1970s atmospheric shift (also detected by Romolo *et al.* (2006) for the upper reaches of the Peace River Basin). The Mackenzie River basin, lower St. Lawrence River region, and a broad band stretching across northern Ontario and the Prairies were affected greatest by this downward trend in snow accumulation (Brown and Braaten, 1997). Significant decreases were also noted in spring snow cover extending across the western half of the country, as well as in summer snow cover over much of the Arctic. Similar observations were noted in a hybrid study of snow depth and extent of snow coverage over North America from 1960 to 2000 (Dyer and Mote, 2006). Though little change was detected between November and January, with the exception of localized decreases in central Quebec and the Mackenzie River basin, widespread decreasing trends in snow depth were present throughout the continent in February and particularly March. The greatest decreases in snow depth occurred in central Canada, between the Yukon Territory and the Great Lakes region. Mote *et al.* (2005) similarly found widespread declines in springtime

SWE over much of western North America over the period of 1925-2000, especially since mid-century. Dyer and Mote (2007) related these significant decreases in spring snow to a positive trend in the frequency of snow ablation in March and April.

2.1.2 Hydrology

The implications of changing surface climate stretch beyond the scope of meteorological variability. Changing precipitation patterns, increased temperatures, reduced snow cover, and melting of snow and glacial ice have all been linked to changes in regional hydrology (Bates *et al.*, 2008). A recent report of the IPCC (Bates *et al.*, 2008) stressed that future climate is likely to intensify the hydrologic cycle and hence may affect water security in certain locations. For some regions, this will mean enhanced access to water resources, but because the effects will not be spatially uniform, other locations will experience reduced access (Prowse, 2008). Water resources are essential for municipal, industrial and agricultural use, as well as hydroelectricity generation and ecosystem integrity, hence understanding the spatial redistribution of runoff is of utmost importance.

Despite limitations in data, regional storage effects, and competing influences of precipitation and temperature (evaporation), all of which can cause differences in the runoff response of a watershed (Prowse, 2008), similarities among climate and runoff have been observed in regions that spatially align with noted increases in temperature and precipitation – namely, the northern high latitudes. In addition, global scale studies using hydrologic and general circulation models (GCMs) have predicted future changes in water distribution that will see heightened runoff (>20%) at high northern latitudes and decreased runoff in the mid-latitudes (Barnett *et al.*, 2005; Bates *et al.*, 2008). Most analyses, both past and modelled future, show strong gradients within North America, and particularly in western Canada.

2.1.2.1 Western Canadian Water Availability

Water availability over the majority of western Canada is controlled by snowmelt from the Rocky Mountains. In a nival, or snowmelt-dominated, regime, spring snowmelt forms the major flow events of the year (Prowse and Ommanney, 1990). Nival rivers exist in both the Arctic and subarctic, where the latter generally experience larger discharges due to contributions of spring rainstorms. In some rivers, peak flows do not cease after spring snowmelt, but continue to rise throughout the summer due to contributions from glacial runoff; this type of hydrological regime is known as proglacial. Changes in both temperature and precipitation can drastically affect runoff in both nival and proglacial rivers. As cold season temperatures rise, less precipitation tends to fall as snow, resulting in a shallower snowpack and associated reductions in runoff. Warmer spring temperatures have also been linked to earlier spring melt, and consequentially reduced flows during the summer and fall, when water demand is highest (Barnett *et al.*, 2005). Due to the prominent relationship between snow depth/cover and river runoff, changes in snow patterns may intensify variability in the hydrologic regime in snow-dominated basins.

Mountain watersheds are key sources of water for downstream users (Kienzle *et al.*, 2012), especially in western Canada. A number of recent studies have conducted trend analyses on runoff from rivers originating in the snow-dominated Rocky Mountains. Stewart *et al.* (2004) found widespread and regionally coherent trends toward earlier onset of spring snowmelt by 1-4 weeks over most of western North America, and attributed this shift to a combination of effects from the PDO and a long-term springtime warming trend spanning both phases of the PDO. St. Jacques *et al.* (2010) found that, even by eliminating the effect of the PDO, a major controlling factor of streamflow in the northern Rockies, overall runoff in this mountainous region has still declined over the past 60 years. Whitfield and Cannon (2000) similarly identified increases in winter flows, earlier timing of spring peak flows, and substantial decreases in summer month discharge in northern BC and the Rocky Mountains; these trends were attributed to warmer year-round temperatures and increased precipitation in early winter. Rood *et al.* (2008) noted that the greatest changes in Rocky Mountain river flows occurred in rivers draining the east-slopes of the Rocky Mountains towards the northern Prairies and Hudson Bay, with late summer flows declining at a rate of about 0.2%/year. For the alpine-fed Liard and Athabasca River basins, Burn *et al.* (2004b) identified a general trend to earlier onset of spring runoff; increasing winter flows and earlier occurrence of the spring freshet were similarly identified for the larger Mackenzie River Basin (MRB) by Abdul Aziz and Burn (2006) and Burn (2008).

In the Canadian Arctic, Déry and Wood (2005) found generally decreasing annual discharge in a collection northern Canadian rivers, including several feeding into the Arctic Ocean from the western territories; they related these observations to changes in precipitation rather than evapotranspiration, as well as to various large-scale teleconnections, including the AO, ENSO, PDO, and Pacific North American (PNA) atmospheric mode. In northern BC and the Yukon, Fleming and Clarke (2003) examined trends in annual streamflow and determined that glacial cover is a major controlling factor of the fluvial response to climate change; specifically, annual flow was shown to have increased in glacier-fed rivers but contrarily decreased in nival steams. Fleming *et al.* (2006) related this difference to contrasting relationships between watersheds with and without a glacial cover and the AO: the annual glacial signal was found to be significantly positively correlated to the AO index, whereas the nival signal was not. For the same region, Whitfield (2001) detected overall increased streamflow throughout the majority of the year, particularly in the winter months, due to a warming of winter temperatures and consequential release of stored water from glaciers, aufeis, permafrost, and snow deposits. Whitfield and Cannon (2000) similarly found year-round increases in discharge caused by warmer and wetter conditions. Zhang *et al.* (2001) noted increases in annual daily mean streamflow in this region.

For interior BC, Déry *et al.* (2009) indicated that an earlier onset of the spring melt, in addition to decreases in summer streamflow and a delay in the onset of fall discharge, are common to nival and glacial, as well as rainfall-dominated, rivers in this region. Whitfield (2001) found that streams in this region displayed a common shift toward an earlier spring freshet, in addition to reduced overall streamflow as a result of extended summer warm temperatures and reduced fall precipitation. Zhang *et al.* (2001) similarly identified significantly earlier onset of the spring melt for both 40- and 50-year study periods, and extended this observation to all of BC.

In southern BC, many streams are no longer pristine due to extensive changes in land-use, mining, and water withdrawals. Whitfield and Cannnon (2000) identified an earlier onset of spring runoff, followed by lower flows throughout the late summer and fall. Zhang *et al.* (2001) found increases in March/April discharge, but contrasting significant decreases in mean annual streamflow. Significant decreases in maximum and minimum daily mean streamflows were also observed. Whitfield (2001) noted overall decreased discharge in this region, particularly during the winter months.

Reductions in mountain snowpack and changes in snowmelt-derived streamflow timing from higher elevations are also of considerable concern to downstream arid regions where human water demand already equals or exceeds renewable water supply. For many arid and semi-arid lowland regions, such as the Canadian Prairies, surface water is derived from precipitation falling as rain or snow in nearby higher-elevation watersheds. While rain immediately contributes to streamflow, mountain snowpack serves as a natural reservoir for cold-season precipitation storage, releasing water during the warmer months to the drier and much hotter valleys below (Stewart, 2009). In the dry, Canadian Prairies, recent trends and future projections include lower summer streamflows, falling lake levels, retreating glaciers, and increasing soil- and surface-water deficits (Sauchyn and Kulshreshtha, 2007). In rivers feeding the western Prairies, Schindler and Donahue (2006) found drastic reductions in summer discharge since the early 20th century, ranging from 20% in rivers where human impact has been minimal (such as the Athabasca) to 84% in regions where damming and large water withdrawals have occurred (including the South Saskatchewan River). Comeau *et al.* (2009) studied the input of glacier melt to the North and South Saskatchewan Rivers and projected that the greatest impact of glacier decline will be manifested in the timing of flow as opposed to annual volume, with a decline in late summer streamflow. Fleming and Sauchyn (2013) examined the paleohydrology of the Saskatchewan River Basin (SRB), and found that water supply in this region varies on a generational-scale, and that hydroclimatic relationships between adjacent north and south sub-basins, and with the PDO, are profoundly nonstationary. For the South SRB, Tanzeeba and Gan (2012) used four IPCC emissions scenarios and four GCMs to model future climate and water availability; their results suggested that enhanced evaporation caused by rising temperatures will offset increases in precipitation, contributing to reduced mean annual maximum flows. Kienzle *et al.* (2011) similarly simulated a range of climate change conditions for the upper North SRB, and found that, under all climate change scenarios, a shift in the future hydrologic regime will include earlier spring runoff and peak flows, increases in both high and low flow magnitudes, and significantly greater discharge between October and June, with correspondingly less between July to September. Barnett *et al.* (2005) indicated that an increase in surface air temperature in the Canadian Prairies will result in a decrease in winter snow pack, earlier snowmelt, and a decrease in summer soil moisture; these effects, combined with a longer period of low flows during summer and fall, could lead to an increase in the frequency and severity of droughts (see also Bonsal *et al.*, 2013). Wolfe *et al.* (2011) warned that regions dependent on high elevation runoff must prepare to cope with impending water scarcities, especially given that glaciers are smaller now than at any time in the last 3,000 years (Luckman, 1998).

The hydrograph of rivers along the west coast of BC differs from the majority of western Canada in that it follows a pluvial regime. This region is warmed by the oceans and hence does not experience sufficient snowmelt to be classified as nival or proglacial, instead experiencing heavy rainfall throughout the year. Nevertheless, the Pacific coast basins are a critical supplier of water resources in western Canada. Whitfield (2001) examined a group of coastal watersheds in BC and found reduced streamflows year-round, especially during the fall, due to a lengthening of the dry season. Whitfield and Cannon (2000) similarly identified generally lower summer flows, but also detected higher winter flows and a delayed spring peak that they attributed to a shift towards warmer winter temperatures and higher winter precipitation.

2.2 Overview of Data and Methods

Having outlined the key changes in western Canadian hydroclimatology over the past century, this section presents an overview of the data and methods frequently applied in hydroclimatic research, and in particular, for quantifying and classifying water availability and change.

2.2.1 Quantifying Water Availability

Water availability can be measured using a number of different variables and techniques. In smaller watersheds, a water balance equation can be used to quantify local inputs from precipitation and groundwater and outputs to streamflow, evapotranspiration and groundwater recharge (Thonthwaite and Mather, 1955; Newton, 2013). In larger, snow-dominated regions, SWE can be used to represent water availability as frozen storage. Although SWE can be calculated using a variety of different methods, including in-situ gauge measurement or remote sensing (Derksen and LeDrew, 2000), spatial and temporal variability of SWE can be large due to differences in climate, land use, forest cover, and topographic variation (Goita *et al.*, 2003). Techniques are therefore generally not suitable for basins lacking surface homogeneity or for comparison among basins of different terrains (e.g., mountain versus prairieland) (Newton, 2013).

Climate stations are used to continuously measure and collect data on precipitation, temperature, and other climate variables pertinent to understanding weather conditions at a given location. Although the spatial extent and length of records for climate stations is favourable in comparison to SWE measurement data, climate station data do also contain some inherent downfalls. Precipitation measurements do not account for sublimation, which may cause considerable loss to snowpack. MacDonald *et al.* (2010), for example, simulated alpine blowing snow and snowpack sublimation in the southern Rocky Mountains and found that snow mass losses due to sublimation were significant, ranging from 20-32% of cumulative snowfall. Goulding (1978) noted that sublimation rates during Chinook winds resulted in significant moisture loss on the eastern slopes of the Rocky Mountains, with potential evaporation averaging 1.2 mm/day in 1975 and 2.0 mm/day in 1976.

The spatial distribution and operation of climate stations also present some challenges and inconsistencies. Climate stations are often located near populated areas, and therefore may not adequately capture meteorological characteristics of other, more remote locations (e.g., at the tops of mountains). In addition, while a network of hundreds of stations exists for southern Canada, including some with data records extending back 100 years, northern Canada contains few long-term climate stations – just 5% of the available Canadian station network is located north of a line defined by latitude = -0.15 × longitude + 42.0 (Hutchinson *et al.*, 2009). The resulting uncertainty regarding observed climate in the North is further increased by measurement biases, such as gauge undercatch, which can be greater than 50% for frozen precipitation in windy environments (Goodison *et al.*, 1998).

Interpolation of climate data between stations can be used to create gridded data sets covering larger spatial extents; a number of different methods exist for accomplishing this task. Zhang *et al.* (2000) described a procedure for creating a 50 km resolution temperature and precipitation data set for Canada based on the method developed by Hogg *et al.* (1997). This method was applied in the development of Environment Canada's CANGRID climate data product, which has been used in a number of more recent studies (e.g., MacKay *et al.*, 2003; Bonsal and Kochtubajda, 2009; Peters *et al.*, 2012). Bonsal and Prowse (2006) found overall consistency among temperature and precipitation data from four different gridded data sets, including the Climatic Research Unit (CRU, 0.5° resolution), the Inverse Distance Weighting (IDW, 50 km resolution), the square-grid (1 arc minute resolution), and the Australian National University Spline Interpolator (ANUSPLIN, 5 arc minute resolution) global climate data sets. McKenney *et al.* (2011) created a 10 km grid resolution climate data set for Canada using the ANUSPLIN method of thin-plate smoothing splines. The ANUSPLIN interpolation method takes into account topography in addition to a variety of other input parameters when defining input points (Newton, 2013). Temperature and precipitation estimate errors were determined to be similar to instrument measurement errors, hence the quality of the data set is assumed to be robust (McKenney *et al.*, 2011). Hutchinson *et al.* (2009) assessed the ANUSPLIN model by withholding 50 climate stations in southern Canada from the gridding procedure, and then compared the omitted records with interpolated values; modelled and observed data were found to be similar with minimal biases. Withheld validation errors were also calculated for northern Canada, and while errors were significantly larger, they were not as large as would be expected given that the data in the North are ~20 times sparser than in the south. It was suggested to use the northern surfaces with moderate confidence.

Similar to climate data collection stations, hydrometric data, including daily and monthly mean streamflow, water level, peaks and extremes, and sediment concentrations, are recorded at over 2,500 active hydrometric monitoring stations across Canada forming part of the Water Survey of Canada (WSC) (Environment Canada, 2013b). Monitoring technology is currently a mix of ageing analogue water level recorders and modern digital recorders. Hydrometric stations are located on lakes, rivers, and streams of many sizes, ranging from drainage basins as small as a few hectares to watersheds larger than a million squared kilometres, and like climate stations, are far more heavily concentrated in the southern half of the country where population and economic pressures are greatest. As a result, the adequacy of the network to describe hydrologic characteristics, both spatially and temporally, decreases significantly to the north (Environment Canada, 2013a).

The Reference Hydrometric Basin Network (RHBN) is a sub-set of the national network that has been identified for use in the detection, monitoring, and assessment of climate change. The criteria according to which 255 stations were selected for membership in the RHBN include (Harvey *et al.*, 1999):

1. *Degree of basin development*. Only stations reflecting pristine catchments are included in the network. As a guideline, a catchment with less than 10% of its surface area modified from natural conditions is considered pristine.
2. *Absence of significant regulations or diversions*. A catchment is considered natural (unregulated) if there is no control structure upstream of the gauging station.
3. *Record length*. A minimum of 20 years of data is required for a station to be included in the RHBN.
4. *Longevity*. A station is excluded from the network if it is currently active but is not expected to have future data collection activities (as determined by regional staff).
5. *Data accuracy*. Only stations with good quality data, as assessed by local experts based on knowledge of the hydraulic condition of the stations, are included in the RHBN.

When assessing impacts of climate change on hydrology, it is important to choose locations where there has been minimal development or human impact on the basin (i.e., stations forming part of the RHBN). Unfortunately, especially given the dense populations and extent of development in southern Canada, it is not always possible to find pristine basins. It is therefore important to be mindful of other influences, such as regulation or extraction, when linking changes in water resources to changes in climate.

2.2.2 Trend Testing with Hydroclimatic Time Series

The detection of hydrologic and climatic trends may be sought in historical records providing such data are representative and cover sufficient periods of time (Zhang *et al.*, 2000). A number of different methods have been used for the detection of trends in hydroclimatic research, including the Mann-Whitney or Wilcoxon rank-sum test (Gobena and Gan, 2006; Fleming and Sauchyn, 2013), the two-sided difference of means test (Dyer and Mote, 2006), and the least squares line of best fit (Hidalgo *et al.*, 2009), in addition to parametric methods (Jonsdottir *et al.*, 2008). Arguably the most widely used method, however, is the Mann-Kendall (MK) non-parametric test for trend (e.g., Burn *et al.*, 2004a,b; Déry and Wood, 2005; Burn, 2008; Burn *et al.*, 2008). Developed by Mann (1945) and Kendall (1975), the MK test was found to be an excellent tool for statistical trend detection by numerous researchers (Burn and Hag Elnur, 2002;

Booth *et al.*, 2012; Déry *et al.*, 2012; Peters *et al.*, 2013). *Section 4.2.1.2* discusses the MK test in depth.

2.2.2.1 Assessing Trend Results

Certain factors can undermine the validity of hydroclimatic trend results. Short data records can lead to Type 1 error – that is, rejecting the null hypothesis when it is actually true – as anomalous events may be unrealistically magnified in otherwise normal conditions (Li, 2012). Choosing a data set of sufficient length helps to minimize this problem, as statistical confidence is increased by increasing degrees of freedom. It is recommended to use time series with at least 30 data points.

Change-points in a data set can also lead to false positives in a trend test, hence assessing the relative homogeneity of the data prior to studying trend is recommended. Change-points, or regime shifts, are times of discontinuity that can be attributed to a number of different factors, including changes in observation location, equipment, measurement techniques, or environmental or atmospheric conditions (Reeves *et al.*, 2007). Continuity, or a lack of change-points, is often necessary to draw accurate conclusions from trend analyses. Change points in hydrologic or climatic time series can be detected using a number of different statistical methods, including the Pettitt non-parametric test, the standard normal homogeneity test, the Buisand range test, the von Neumann ratio test, Wilcoxon's non-parametric test, two-phase regression procedures, inhomogeneity tests, information criteria procedures, Bayesian change point analysis, and various variants thereof (Reeves *et al.*, 2007; Ming Kang and Yusof, 2012).

The presence of serial correlation can similarly complicate the identification of trend in a data set (Burn and Hag Elnur, 2002). Positive serial correlation, often present in hydrological and meteorological time series, can increase the expected number of false positive outcomes detected by the MK test (von Storch and Navarra, 1995; Zhang and Zwiers, 2004). The problem of serial correlation, and hence of Type 1 error, is commonly dealt with by pre-whitening the data prior to running trend analysis. Pre-whitening involves estimating and removing the lag-1 autocorrelation $\widehat{\alpha}$ from the series and replacing the original time series X_t by:

$$Y_t = X_t - \hat{\alpha} X_{t-1} \tag{2.1}$$

Fleming and Clarke (2003) argued that pre-whitening can substantially and inappropriately decrease the power of the trend significance and increase error in slope estimates, but noted that multi-stage pre-whitening techniques appear to produce superior results. Yue *et al.* (2002) similarly demonstrated that the common pre-whitening procedure can lead to potentially inaccurate assessments of the significance of trend, and recommended a modified method in which a trend first be removed from the series prior to ascertaining the magnitude of serial correlation. Bayazit and Onez (2007) concluded that the pre-whitening procedure need not be applied when dealing with large samples ($n \geq 50$) and high magnitude trend slopes, or when the coefficient of variation is very small; in other cases, however, pre-whitening will prevent the detection of a non-existent trend and should therefore be used.

Cross-correlation in the data can lead to an increased number of expected trends under the hypothesis of no trend in the data (Lettenmaier *et al.*, 1994; Burn and Hag Elnur, 2002). It is therefore important to consider the field, or global, significance of the trends when determining whether the number of observed significant outcomes is unusual. Field significance reveals the percentage of tests that are expected to show a trend at a given local significance level. The conventional approach for evaluating field significance in hydrological and climatological studies is to count the number of individual (or "local") tests yielding nominally significant results and then to judge the unusualness of this integer value in the context of the distribution of such counts that would occur if all local null hypotheses were true (Wilks, 2006). Other methods for calculating field significance include Walker's test of minimum p-values, the global test based on the False Discovery Rate (Wilks, 2006), and a bootstrap resampling approach (Burn and Hag Elnur, 2002; Burn *et al.*, 2004a,b).

2.2.2.2 Interpreting Trend Results

While non-parametric trends tests provide an indication of the presence, direction, and significance of a trend, it is difficult to assess or compare the trend (against, for example, another trend) without knowing its magnitude. The robust trend slope estimator developed by Sen (1968) is frequently used to quantify the magnitude and direction of trends (Burn and Hag Elnur, 2002; Burn *et al.*, 2008).

2.2.3 Classifying Regions of Coherent Behaviour

In hydroclimatological applications, it is often desirable to delineate groups or classes of coherent hydrologic or climatic behaviour to classify regions according to some property; this is frequently known as regionalization. Classification or regionalization involves reducing a large amount of data into a manageable set of patterns that can be used to facilitate analysis. There are a number of ways to accomplish this task, including both statistical and non-statistical methods. Having *a priori* knowledge, researchers can select non-hydrological (or non-climatic) variables known to be of importance in depicting flow (or climate) regimes (Li, 2012). Snelder *et al.* (2005), for example, used "controlling factors" known to be responsible for causing physical and biological variability in rivers, including climate and topography, to classify flow regimes in New Zealand. Whitfield (2001) classified groups of BC rivers by visual examination of the hydrographs (rainfall-runoff, snowmelt, etc.) and by the apparent timing shifts in specific time periods. Haines *et al.* (1988) used monthly mean flows to produce a global map of flow regimes.

Eigentechniques have been used extensively in the meteorological, and more recently, hydrological communities for data reduction, grouping of variables, and identification of coherent modes (Piechota *et al.*, 1997). One common technique to accomplish this is Principal Components Analysis (PCA) (also known as Empirical Orthogonal Function (EOF) analysis). PCA has been used widely for its success in delineating regional patterns of precipitation, temperature, sea level pressure, atmospheric circulation, drought, cyclone frequency, and streamflow/runoff (refer to Lins, 1997 for a list of references). Lins (1997), for example, used PCA to define the dominant regions of interannual streamflow variability in the United States. Cayan *et al.* (2001) used EOF analysis to delineate patterns of lilac first bloom dates (a measure of the onset of spring) in the western United States. Maurer *et al.* (2004) used PCA to identify spatial patterns of runoff using a gridded seasonal data set to analyze sources of runoff predictability. Chen *et al.* (2009) used EOF analysis to map precipitation regimes in China.

Other statistical methods for classifying hydrologic or climatic patterns include cluster analysis (CA) and self-organizing maps (SOM). Among conventional CA techniques, K-means partitioning method, average linkage, complete linkage, and Ward's method have been widely used with convincing results (Haines *et al.*, 1988; Piechota *et al.*, 1997; Stahl and Demuth, 1999). In a study of the western United States, McCabe (1996) used a hierarchal average-linkage clustering method to identify groups of stream gauges with intercorrelated annual streamflow to facilitate correlation analysis between streamflow and winter mean 700-hPa height anomalies. Li (2012) used Ward's agglomerative method to classify streamflow regions in New Zealand. Kalteh *et al.* (2008) provided a review of the SOM algorithm and its application in water resources problems. The windowed Fourier transform or continuous wavelet transform can also be used to describe spatial and temporal variability of seasonal streamflow (Coulibaly and Burn, 2005).

2.2.4 Linking Hydrology and Climate

As reflected in past studies (*Section 2.1.2*), climate has significant influence on both local and regional hydrological patterns. This section identifies methods for statistically linking trends and patterns in streamflow/runoff with both surface climate and atmospheric patterns.

2.2.4.1 Correlation Analysis

The relationship between hydrological activity (e.g., streamflow or runoff, timing of a hydrologic event) and climate change (e.g., temperature, precipitation) is frequently measured via direct correlation analysis (Cayan *et al.*, 2001; Burn *et al.*, 2004a; Ahmed *et al.*, 2013). In statistics, the Pearson product-moment correlation coefficient, or Pearson's *R*, is a measure of the linear correlation (dependence) between two variables X and Y, and is a value between +1 and −1 inclusive. It is widely used in the sciences as a measure of the strength of linear dependence between two variables. In addition to meteorological variables, this method has also been extended to measure teleconnections between large-scale atmospheric patterns and hydrological variables (Burn *et al.*, 2004a; Déry and Wood, 2005).

A variation of direct correlation analysis, partial correlation analysis has also been used to determine the strength of relationships between hydrological and meteorological variables (Burn *et al.*, 2004b; Burn, 2008). Partial correlation analysis involves calculating correlation between hydrological variables that exhibit a significant trend and meteorological variables in which a significant trend has been detected. Partial correlations allow for the identification of correlation between variables independent of any common trend signal in the two variables, thus making it possible to attribute observed trends in hydrological measures to climate change (i.e., trends in meteorological variables) (Burn, 2008).

2.2.4.2 Composite Analysis

Linkages between hydrological variables and climate indices representing global atmospheric patterns are frequently examined by means of a composite analysis approach (Maurer *et al.*, 2004; Burn, 2008; Burn *et al.*, 2008). Since climate signals can exhibit strong terrestrial teleconnections in one phase but weak teleconnections in an opposing phase, a direct linear correlation could fail to detect important signals (Maurer *et al.*, 2004). The procedure for composite analysis involves examining high and low values for the climate indices separately by choosing subsets of years representing the largest and smallest values in the series for each climate index (e.g., years corresponding to the largest/smallest 10 climate index values or largest/smallest 25% of climate index values) and compositing the hydrological data for corresponding years for each category. A t-test is routinely performed to determine if the subset of values chosen for the hydrological variables (i.e., the dependent variables) differ significantly from the series mean. This is sometimes known as a "circulation-to-environment" approach (Li, 2012); an "environment-to-circulation" approach involves the opposite – choosing the extreme highest and lowest values in the hydrological variable series and performing a t-test on climate index data for corresponding years. Dependent series should be tested for normality to ensure the power of the t-test. Cayan (2001) extended the composite analysis method to examine the relationship between spring freshet (as measured by the lilac first bloom) variability and meteorological variables, including temperature and precipitation.

A number of atmospheric patterns have been known to influence hydrology and flow metrics in Canada, and could therefore be applied in composite analysis. Déry and Wood (2004) and Fleming *et al.* (2006) examined the relationship between Canadian streamflow and the AO. Déry and Wood (2005) linked trends in northern Canadian river discharge with variability in the AO, ENSO, PDO, and PNA climate patterns. Bonsal *et al.* (2006) investigated the impacts of the ENSO, PDO, PNA, AO, North Pacific (NP) index, and North Atlantic Oscillation (NAO) patterns on Canadian freshwater ice break-up and freeze-up dates. Burn (2008) explored the relationship between various measures of the spring freshet and the ENSO, PDO, AO, NP, NAO, and the Atlantic Multidecadal Oscillation (AMO); Maurer *et al.* (2004) similarly analyzed these climate indices with respect to different potential sources of runoff in North America. Characteristics of the dominant modes of atmospheric circulation over western Canada are presented in *Section 4.1.3*.

2.3 Summary

Western Canadian climate has shown strong warming trends over the past 50 to 100 years, with contrasting north-south changes in precipitation and streamflow patterns. Many different techniques have been applied for the detection and quantification of streamflow trends and spatial patterns, but studies have generally focused on either smaller regions (i.e., individual watersheds) or larger scales (i.e., all of Canada). The following chapters will outline the study region analyzed in this work, as well as how a number of the discussed research methods were applied for the quantification and analysis of changing water resources in western Canada.

Chapter 3

Study Area

As discussed in *Chapter 2*, the impacts of climate change in Canada have been felt most strongly in the western and northern regions of the country. In addition, climate models predict a continuation in the direction of intensifying climate that will be most extreme in these areas. This research therefore focuses on watersheds located in western and northwestern Canada.

Section 3.1 presents an overview of western Canada's climatic zones. Climate has important implications on the hydrology of a region. *Section 3.2* outlines western Canada's major hydrological regions, including a description of the geography and hydrology of the major drainage basins. The CROCWR study region is introduced, including the physical characteristics associated with each watershed and the methods used to delineate them.

3.1 Western Canadian Climate

Western Canada covers a large geographical region that includes many diverse climatic and ecological zones. The region is generally defined to include the provinces of BC, Alberta, Saskatchewan, and Manitoba, as well as the Yukon, Northwest Territories, and Nunavut.

Climatic regions are often classified based on the Köppen-Geiger Climate Classification System (Köppen, 1936). The Köppen-Geiger classification uses six letters to divide the world into six major climate regions: A, B, C, D, E, and H. Each category is based on annual and monthly averages of temperature and precipitation, and is further divided into sub-categories based on these values.

Western Canada embraces five of the Köppen-Geiger major climatic types: dry (B), mild mid-latitude (C), severe mid-latitude (D), polar (E), and highland (H). A map of Canada's Köppen-Geiger climate classifications is given in Figure 3.1, with western Canada highlighted. Table 3.1 lists typical climate characteristics for each of western Canada's climate regions (sub-class acronyms).

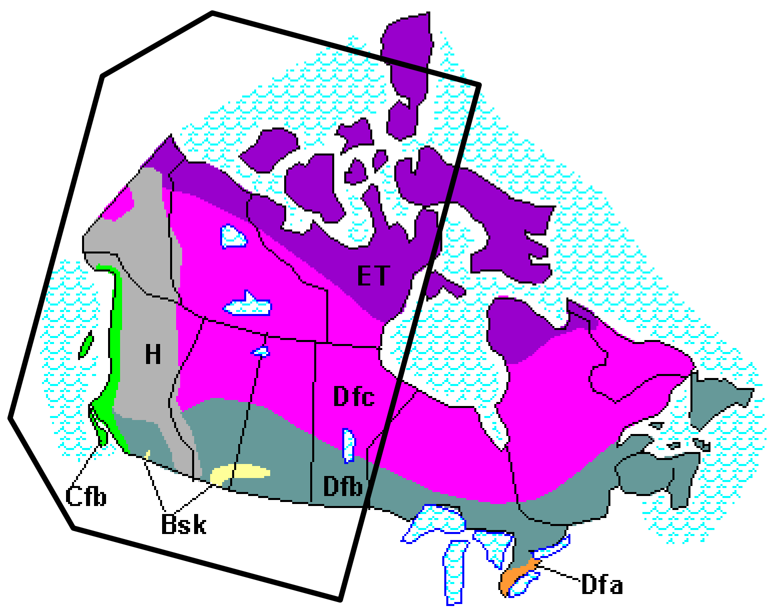


Figure 3.1. Map of Canada's major Köppen-Geiger climatic regions. Western Canada is encompassed within the black box (Climate, 2007).

Embedded within these classes are Canada's major climatic zones: the Prairies, Pacific Maritimes, boreal forest, taiga, Arctic, and Western Cordillera. Each of these zones depends on a number of different conditions, including proximity to large bodies of water, altitude, and latitude.

Table 3.1. Köppen-Geiger climate classification system for western Canada (Climate, 2007).

K-G major class	K-G major class title	K-G sub-class	K-G sub-class title	Sub-class description
B	Dry	Bsk	Mid-latitude steppe	Mid-latitude dry
C	Mild mid-latitude	Cfb	Marine west coast	Mild with no dry season, warm summer
D	Severe mid-latitude	Dfb	Humid continental	Humid with severe winter, no dry season, warm summer
		Dfc	Subarctic	Severe winter, no dry season, cool summer
E	Polar	ET	Tundra	Polar tundra, no true summer
H	Highland			

Dry Prairie Climate (Bsk)

Southern Alberta and Saskatchewan are semi-arid regions characterized by the fact that evaporation is greater than precipitation. The southern Prairies are susceptible to extended periods of drought, but also to periods of extreme precipitation and/or flooding, making interannual variability of water resources high (Shabbar *et al.*, 2011). Climate is characterized by very cold winters and very hot summers (Canadian Geographic, n.d.). Spring showers and temperate autumn weather makes the Prairies one of the top grain-growing areas of the world (Your Canada, n.d.).

Mild Maritime Climate (Cfb)

The Pacific Maritime region lies along the west coast of BC, extending from the Canada-US border in the south to the Alaska border in the north. Hydroclimate in this region is characterized by a pluvial regime and differs greatly from the rest of Canada due to its close proximity to the Pacific Ocean. Warm winds promote a year-round mild, rainy climate, making summers cooler and winters warmer than the rest of the country. Precipitation occurs every month of the year, and there is no discernable dry season, though summer sees less rainfall than winter (Whitfield, 2001; Climate, 2007; Canadian Geographic, n.d.).

Humid Boreal Climate (Dfb)

The southern boreal forest spans a continuous, mid-latitude belt through western Canada, stretching from the Rocky Mountains to Manitoba. The area is dominated by coniferous forests, particularly spruce, interspersed with vast wetlands, mostly bogs and fens. The boreal region exhibits a humid continental climate, with cold winters (average temperature of coldest month < -3°C) and warm summers (hottest month average temperature > 10°C) (Climate, 2007). Convectional precipitation occurs mainly in summer, when land is heated by warmer temperatures (Iu and Richard, n.d.).

Subarctic Taiga Climate (Dfc)

The subarctic, taiga, or northern boreal forest climate zone spans the entirety of western Canada, appearing in every province and territory. The taiga landscape consists mainly of spruce trees. This region is characterized by long, severe winters, where mean temperatures are below freezing for up to six months. Taiga summers are short and cool, and only 50 to 100 days of the year are frost-free. Precipitation in this region is low and occurs mainly in the summer as convectional precipitation (Canadian Geographic, n.d.; Iu and Richard, n.d.).

Polar Arctic Climate (ET)

The high latitudes are dominated by intensely dry and frigid conditions. The Arctic region includes most of the Northwest Territories and Nunavut, and extends into the Yukon cordillera. Landscapes in this vast region range from flat, frozen tundra to mountains and fjords. Summers are very short and cool (warmest month average temperature < 10°C), while winters are long and cold, lasting as long as ten months in the most northerly locations (Iu and Richard, n.d.). The region experiences very low precipitation because most water bodies, including the Arctic Ocean, are frozen for the majority of the year; moreover, much of the precipitation that does occur is not locally derived, but is transported by atmospheric moisture transport and "atmospheric rivers", relatively narrow poleward-moving moisture plumes associated with frontal dynamics (Newman *et al.*, 2012; Zhang *et al.*, 2012). In the extreme high Arctic, snow accumulation is so little that the area is correctly classified as a polar desert (French and

Slaymaker, 2012). While evaporation in the Arctic is generally quite low, losses to sublimation can be significant (Prowse, 2013, personal communication).

Cordilleran Highland Climate (H)

The interiors of BC and the Yukon are dominated by mountain ranges, including the Coast, Columbia, and Rocky Mountain chains. The climate in this area varies greatly due to the presence of mountains. Coastal temperatures tend to be warmer than those inland, and climate in the south is warmer than in northern regions, where winters are long and cold, lasting up to eight months (Whitfield, 2001). Locations only a few kilometres apart can experience very different temperatures and precipitation patterns. As moist air from the coast moves over the mountains, it cools and falls on the western slopes as heavy amounts of rain and snow. The leeward slopes and valleys between the mountain ranges, by contrast, experience hot summers and are almost completely devoid of precipitation due to rain shadow conditions (Iu and Richard, n.d.). The region overall is characterized by a nival regime, and experiences the highest snowfalls in western Canada (French and Slaymaker, 2012).

3.2 Western Canadian Hydrology

Western Canadian rivers drain to four major bodies of water: the Arctic Ocean, the Pacific Ocean, Hudson Bay, and the Gulf of Mexico. Figure 3.2a provides a map of Canada's major drainage basins. The CROCWR study region contains 25 sub-watersheds that drain to three of these bodies: 16 are located in the greater Mackenzie River basin and flow to the Arctic Ocean; five contain rivers that discharge into the Pacific Ocean; and four are located in the Saskatchewan River basin (SRB) and drain to Hudson Bay. Figure 3.2b presents the CROCWR study area, divided (by colour) into major drainage basins. The majority of the CROCWR watersheds are sourced by the Rocky Mountains, which feed the primary tributaries of the Mackenzie River, Canada's largest freshwater contribution to the Arctic Ocean, and the Saskatchewan River, the primary water resource for the Prairie provinces (Newton, 2013). The Fraser and Columbia Rivers are also sourced by the Rockies, and flow westward to the Pacific Ocean. Pacific coast rivers are sourced by glaciers and lakes in the Coast Mountains and discharge to the Pacific Ocean.

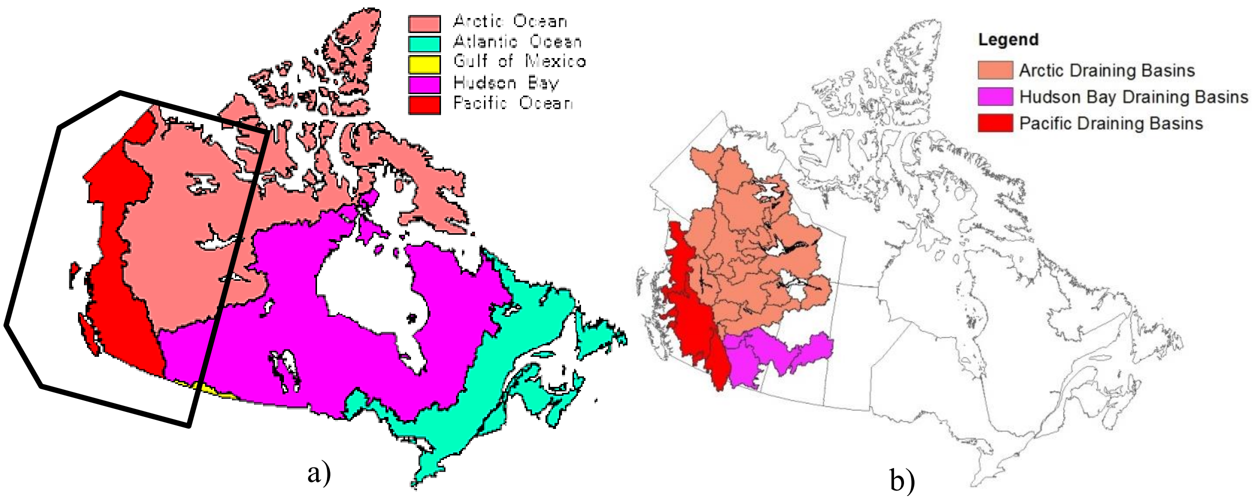


Figure 3.2. a) Map of Canada's five major drainage basins. Western Canada is encompassed within the black box. b) Map of Canada showing the CROCWR study region defined by major drainage path.

Mackenzie River Basin

The Mackenzie River basin extends from central Alberta in the south to the Beaufort Sea coast in the north, and from the continental divide of the western cordillera to the Canadian Shield at the eastern border of the Northwest Territories. Since the basin straddles several climatic regions, annual precipitation varies widely across the region, ranging from ~200 mm/yr at the Arctic coast to over 1000 mm/yr in the southwest. Snowfall is the major form of precipitation, and persists for over half the year in some locations (Woo and Thorne, 2003).

The Mackenzie River basin drains an area of 1.8 million km^2, making it the largest Arctic-draining river in North America. Volumetrically, most of the flow on the Mackenzie River originates outside of the Arctic as mountain snowpack feeding the Liard, Peace, and Athabasca Rivers. The release of freshwater to the ocean creates a thermohaline gradient that prevents extrusion to the denser saline sea water, preserving the integrity of the polar ice pack. Near the outlet, however, ice breakup is amplified by massive spring discharge.

Rivers of the Mackenzie River basin exhibit several seasonal flow patterns, including nival (snowmelt dominated), proglacial (influenced by glacier melt), wetland/muskeg (low relief characterized by poor drainage (Prowse and Ommanney, 1990)), prolacustrine (below large lakes), and regulated flow regimes. Regulation in the Mackenzie River basin has occurred on a number of its tributaries, including the Peace and Arctic Red Rivers, which has caused changes to the hydrographs near the points of regulation, including decreased peak flows and increased winter flows. The effects of regulation, however, become lost in downstream flows due to natural storage effects of the basin, such as that of Great Slave Lake, in addition to contributions from natural (nival or otherwise) tributary streams downstream. In general, streamflow in the Mackenzie River basin is controlled by snowmelt in the spring, as well as convectional and frontal rainfall during summer and autumn.

The mountainous sub-basins in the western Mackenzie River basin, including the Liard and Peace river basins, contribute about 60% of the flow to the Mackenzie, while the interior plains and eastern Canadian Shield contribute only about 25%, despite having a similar total contributing area (~40% of the total MRB area each). The mountain zone is the dominant flow contributor to the Mackenzie in both high- and low-flow years (Woo and Thorne, 2003).

The unregulated Liard River flows north from its Rocky Mountain headwaters, joining the Mackenzie River mainstem at Fort Simpson, Northwest Territories, and exerts the largest influence on streamflow in the Mackenzie River. The flow regime of this subarctic nival river is characterized by high spring flows from snowmelt and river ice breakup, followed by declining summer discharge, and low winter streamflow. Precipitation in the mountainous Liard basin is significantly correlated with altitude and latitude, increasing with proximity to the Pacific Ocean and decreasing to the northeast (Woo and Thorne, 2006).

The Peace River originates in the northern Rocky Mountains and flows northeast across northern Alberta, draining into Lake Athabasca at the Peace-Athabasca Delta (PAD), a UNESCO World Heritage Site and important breeding habitat for a variety of wildlife (Wolfe *et al.*, 2005). The headwaters of the Peace River have been regulated by the W.A.C. Bennett hydroelectric dam since 1968. Peters and Prowse (2001) found that regulation of the upper Peace River has considerably altered the hydrologic regime of the Peace River even some 1,100 km downstream, resulting in increased winter flows, reduced peak flows, and decreased variability in daily flows. Despite reduced peak flows and variability, however, the downstream Peace River hydrograph has still retained the basic shape of the pre-regulation hydrograph, due to increased effects of flow contributions from downstream tributaries.

The Athabasca River originates in the Columbia Icefield near Jasper and empties into Lake Athabasca, serving as both a municipal and industrial water source along its course. The river's proglacial flow regime includes intense glacial ablation in summer, resulting in prolonged spring peak flows (Woo and Thorne, 2003). Summer streamflow in the Athabasca River has declined by 30% since 1970, due in part to natural changes, but mainly caused by the growth of the oil and gas industry (Schindler and Donahue, 2006). The Athabasca River Basin is home to the world's third largest crude oil deposit (Alberta Energy, 2013), the majority of whose developments are located in the lower portion of the Athabasca River. Current mining operations typically consume two to four times as much freshwater as oil produced (Alberta Environment, 2013b), a number that has been forecasted to increase by between 120% to 165% by 2025 (Mannix *et al.*, 2010). In 2010, 74.5% of total Athabasca River surface water allocations were licensed for the oil and gas industry (Alberta Environment, 2013a).

The Slave River links Lake Athabasca to Great Slave Lake, thus connecting the Peace and Athabasca Rivers to the Mackenzie River. Other significant rivers located in the MRB include the Hay, Peel, and Mackenzie mainstem. Major lakes include Lake Athabasca, Great Slave Lake, and Great Bear Lake.

Pacific Coast Basins

A number of both large and smaller rivers along the Pacific coast of BC drain westward from the Coast Mountains into the Pacific Ocean. These rivers are both wild and commercially valuable, particularly within the salmon fishing industry. In the northern region, the Taku, Stikine, Iskut, Nass, Skeena, and Kitimat Rivers flow west, some of which cross into Alaska before emptying into the ocean; these rivers form the North Pacific region. The Wannock, Atnarko, Bella Coola, Dean, and Homathko Rivers constitute the greater South Pacific region. Pacific rivers are characterized by a pluvial regime, in that their hydrology is largely dominated by heavy precipitation, especially during the winter wet season; significantly lower streamflows occur during the summer dry season (Whitfield, 2001).

Fraser and Columbia River Basins

The Fraser River Basin (FRB) is the largest Canadian river by area to drain into the Pacific. The river mainstem, along with its vast tributary network, is located between the Coast and Rocky Mountain ranges in BC. Originating in the Rocky Mountains at the Alberta/BC border, the Fraser River travels northwest through the Rocky Mountain Trench toward Prince George, before turning south at the Nechako River confluence. After passing through the Fraser Canyon, the river turns westward into the Strait of Georgia near Vancouver, and finally empties into the Pacific Ocean (Déry *et al.*, 2012).

Both terrain and climate vary considerably in the FRB. The interior plateau experiences relatively dry conditions due to its position in the rain shadow of the Coast Mountains, whereas the coastal and mountainous sections of the basin are accustomed to heavy precipitation. Snowfall is the dominant component of annual precipitation, especially in the high latitudes and altitudes, and 1.4% of the basin is glacierized. Flow in the FRB is mostly unregulated, with the exception of the Nechako River whose flow has been controlled by the Kenney Dam since the early 1950s (Déry *et al.*, 2012).

The Columbia River is the largest river in the Pacific Northwest region of North America. The river rises in the BC Rockies near Invermere and flows northwest, before turning at the confluence of the Canoe River and heading south into the USA. Approximately 15% of the Columbia River Basin (CRB) is located in Canada, while the remaining area spans seven of the US states (Matheusson *et al.*, 2000).

The Canadian portion of the CRB is located in south-eastern BC and occupies the drainage area upstream of the confluence of the Kootenay and Columbia Rivers. The region experiences a continental climate, and precipitation varies seasonally. The majority of precipitation falls as snow during the winter, and 4% of the basin is glacierized. The Canadian CRB experiences a natural glacial-nival regime in which the hydrologic cycle is dominated by the spring freshet and lowest flows occur in winter (Schnorbus *et al.*, 2012).

The Okanagan River drains the Okanagan region of BC, flowing south across the international border, and joins the Columbia River as a small tributary at the city of Portland. The river rises at the southern end of Lake Okanagan in southern BC and flows through five other lakes before heading into the USA. Climate in the Okanagan River Basin (ORB) is arid and variable. Water in the ORB is heavily allocated, and per person daily water use exceeds twice the Canadian average (Okanagan Water Supply & Demand Project, 2011). ORB streams have consequently experienced reduced discharge, particularly during winter (Whitfield, 2001).

Saskatchewan River Basin

The SRB is one of Canada's largest and most vulnerable inter-provincial watersheds. The SRB extends from the continental divide to Lake Winnipeg, Manitoba, and is a major tributary to the Nelson River, the largest contributor of discharge to Hudson Bay (Déry *et al.*, 2005). Originating in the Alberta Rockies, the North and South Saskatchewan Rivers converge approximately 40 km east of Prince Albert, Saskatchewan to form the greater Saskatchewan River. The South Saskatchewan River is formed at the confluence of the Bow and Oldman rivers near Grassy Lake, Alberta; the Red Deer joins downstream at the Alberta-Saskatchewan border. The North Saskatchewan River itself flows east from the Rockies and is joined by several smaller contributing rivers before reaching the Saskatchewan River Forks. The Rocky Mountain headwaters comprise only a very small proportion of the total basin area, but snowmelt from the mountains is the primary source of streamflow in the Saskatchewan River, accounting for approximately 75% of annual discharge, with 60% of this occurring between early May and mid-July (Pentney and Orhn, 2008).

The SRB faces several trans-boundary water management challenges, both inter-provincially and internationally. The 1969 *Master Agreement on Apportionment* agreement between Alberta, Saskatchewan, Manitoba, and the federal government requires that Alberta pass at least 50% of the annual natural flow in the South SRB downstream to Saskatchewan and Manitoba. Alberta's contribution to the downstream provinces can be drawn from any combination of flows from the Red Deer and South Saskatchewan Rivers, but during some dry years, demand can be greater than Alberta's entire allotted share of the flow. In addition, a small portion of the southernmost extent of the South SRB passes into Glacier National Park in the state of Montana, creating international concerns. The *Boundary Waters Treaty* governs the sharing of the waters between Canada and the United States (Pentney and Orhn, 2008).

Since the late 1800s, surface water in the SRB has been heavily allocated. Water licences are provided for a number of different industries, including municipal, thermal, and industrial sectors, however the largest water user in the South SRB by far is the agricultural sector. Although the share of surface water used for the agricultural sector is ~50% in the Prairies, use in the South SRB is particularly high, at 86.5% (Martz *et al.*, 2007), and 74% for irrigation alone (Pentney and Orhn, 2008). By contrast, municipal water use accounts for only 8.7% of allocations in the South SRB. The Saskatchewan River and its tributaries provide municipal water resources for the cities of Red Deer, Calgary, Lethbridge, Medicine Hat, Edmonton, and Saskatoon, as well as many smaller communities and farms. Water from the South Saskatchewan River is also diverted to the Qu'Appelle River through the Qu'Appelle River Dam and into Buffalo Pound Lake to supply water to the cities of Moose Jaw and Regina, as well as to mining operations outside the South SRB (Martz *et al.*, 2007).

Rivers in the SRB have a history of being heavily regulated. The construction of dams and weirs for flood control, agriculture, and hydroelectric generation has altered flows in the South SRB to a considerable extent (Pentney and Ohrn, 2008). Although large flows can still occur, medium-sized flood events are attenuated by reservoir operations and winter low flows only occur when flows cannot be augmented by reservoir releases. Both the timing and magnitude of such "extreme" (i.e., peak or minimum) flows are altered through reservoir operation, but it is again important to note the diminishing role of regulation on downstream flows, as flows are returned to near-natural levels from contributions of other downstream tributary streams. There are a number of dams and reservoirs within the SRB, including the Oldman River Dam, the Dickson Dam on the Red Deer River, and a system of hydroelectric dams, reservoirs, and canals on the Bow River. The largest dam in the basin is the Gardiner Dam south of Saskatoon, whose 1967 completion created the upstream Lake Diefenbaker. With a maximum capacity of 9.4 km^3, Lake Diefenbaker is a multiuse reservoir which provides water for hydroelectric power production, irrigation, mining, recreation, flood control, and municipal use in southwestern Saskatchewan (WSA, 2012).

3.2.1 The CROCWR Study Region

The CROCWR study region consists of 25 sub-basins forming parts of the major drainage regions outlined above. Watersheds were delineated using ArcHydro along with a series of latitude-longitude points that correspond to 37 WSC hydrometric gauging stations. ArcHydro is an object-oriented model capable of establishing a topological network that includes flow direction, connectivity, and upstream/downstream relationships of stream segments using a digital elevation model (DEM), a stream network, and points to distinguish watershed outlets (Linton *et al.*, 2013). River basins were separated into upstream, downstream, and/or other contributing areas. Table 3.2 provides a detailed summary of the WSC gauging stations used to define each watershed, as well as geographical characteristics of each basin. Median basin elevations were calculated using a Canadian DEM. Refer to Appendix A for hypsometric profiles of each of the 25 study basins. Watershed latitude and longitude coordinates are given for the centroid of the watershed. Stations forming part of the RHBN, in addition to stations located on regulated rivers, are noted.

Figure 3.3 shows a map of the CROCWR region by individual watershed, while Figure 3.4 identifies each station by WSC ID.

3.2.1.1 Flow Estimation of Stations

Three of the WSC stations were installed much later than the others and therefore have considerably shorter streamflow records. To elongate each of these streamflow series to facilitate statistical analysis, measurements from a nearby upstream or downstream gauge were used to estimate early year flows for the short-record stations. To account for missing or additional contributing area, the estimated river streamflow was scaled such that:

$$Q_{\mathrm{S}} = Q_{\mathrm{N}} A_{\mathrm{S}} / A_{\mathrm{N}} \tag{3.1}$$

where Q (m^3/s) denotes the streamflow within a drainage area A (m^2), and subscripts S and N identify the short-record station and the nearby station, respectively (Déry *et al.*, 2005). Stations in which this method was applied are identified in Table 3.2 with "+" superscripts.

Table 3.2. Details of the CROCWR study region.

WSC Station					*CROCWR Watersheds*						
Station ID	**Station Name**	**Latitude**	**Longitude**	**Drainage Area (km^2)**	**Watershed Name**		**Definition (based on WSC stations)[%]**	**Latitude***	**Longitude***	**Drainage Area (km^2)**	**Median Basin Elevation (m.a.s.l.)**
Liard River											
10BE001	Liard River at Lower Crossing	59.4125	-126.0972	104,000	**1**	Upper Liard	10BE001	59.9745	-128.5646	104,000	1,145
10CD001 [N]	Muskwa River near Fort Nelson	58.7883	-122.6592	20,300	**2**	Fort Nelson	10CD001	58.1623	-123.6517	20,300	970
10ED002	Liard River near the Mouth	61.7428	-121.2281	275,000	**3**	Lower Liard	10ED002 - 10BE001 - 10CD001	60.0929	-123.7590	150,700	674
Peace River											
07FD002 [R]	Peace River near Taylor	56.1358	-120.6703	101,000	**4**	Upper Peace	07FD002	56.1671	-123.9107	101,000	1,187
07GJ001	Smoky River at Watino	55.7156	-117.6219	50,300	**5**	Smoky River	07GJ001	54.6636	-118.5648	50,300	863
07KC001[R]	Peace River at Peace Point	59.1172	-112.4369	293,000	**6**	Lower Peace	07KC001 - 07FD002 - 07GJ001	57.1369	-116.8128	141,700	627
Athabasca River											
07AD002	Athabasca River at Hinton	53.4231	-117.5706	9,770	**7**	Upper Athabasca	07AD002	52.8788	-117.9716	9,770	1,960
07DA001	Athabasca River below McMurray	56.7806	-111.4000	133,000	**8**	Lower Athabasca	07DA001 - 07AD002	55.2416	-113.4174	123,230	651
07LE002[+ N]	Fond du Lac at Outlet of Black Lake	59.1472	-105.5389	50,700	**9**	East Lake Athabasca	07LE002	58.8810	-105.4714	50,700	426
07JD002[++]	Wabasca River at Highway No. 88	57.8744	-115.3889	35,800	**10**	West Lake Athabasca	07JD002	57.8487	-111.9975	35,800	341
Great Slave Lake											
07OB001 [N]	Hay River near Hay River	60.7447	-115.8597	51,700	**11**	Hay	07OB001	58.7796	-118.6704	51,700	441
07NB001 [R]	Slave River at Fitzgerald	59.8722	-111.5833	606,000							
Fort Providence[#]	Mackenzie River near Fort Providence	61.2608	-117.5447	980,000	**12**	Great Slave	Fort Providence - 07OB001 - 07NB001	62.1204	-112.6223	322,300	336
Mackenzie River											
10GC001 [R]	Mackenzie River at Fort Simpson	61.8686	-121.3569	1,270,000	**13**	Upper Mackenzie	10GC001 - 10ED002 - Fort Providence	61.6100	-119.2350	15,000	252
10KA001 [R]	Mackenzie River at Norman Wells	65.2739	-126.8442	1,594,500	**14**	Mid-Mackenzie	10KA001 - 10GC001	64.6057	-122.1243	324,500	320
10LC014 [R]	Mackenzie River at Arctic Red River	67.4581	-133.7444	1,679,100	**15**	Lower Mackenzie	10LC014 - 10KA001	66.0067	-129.8112	84,600	245

Table 3.2(con't). Details of the CROCWR study region.

WSC Station					*CROCWR Watersheds*						
Station ID	**Station Name**	**Latitude**	**Longitude**	**Drainage Area (km^2)**		**Watershed Name**	**Definition (based on WSC stations)[%]**	**Latitude***	**Longitude***	**Drainage Area (km^2)**	**Median Basin Elevation (m.a.s.l.)**
Peel River											
10MC002 [N]	Peel River above Fort McPherson	67.2361	-134.9075	70,600	**16**	Peel	10MC002	65.4415	-135.4990	70,600	885
Pacific Basins											
08BB005[+++]	Taku River near Juno	58.5386	-133.7000	16,700							
08CE001	Stikine River at Telegraph Creek	57.9008	-131.1544	29,000							
08CG001[N]	Iskut River below Johnson River	56.7389	-131.6736	9,350							
08DB001	Nass River above Shumal Creek	55.2639	-129.0861	18,400							
08EF001	Skeena River at Usk	54.6306	-128.4319	42,300							
08FF001	Kitimat River below Hirsch Creek	54.0594	-128.6747	1,990	**17**	North Pacific	08BB005 + 08CE001 + 08CG001 + 08DB001 + 08EF001 + 08FF001	56.6472	-129.0422	117,740	1,187
08FA002	Wannock River at Outlet of Owikeno Lake	51.6792	-127.1792	3,900							
08FB006 [N]	Atnarko River near the Mouth	52.3603	-126.0053	2,430							
08FB007	Bella Coola River above Burnt Bridge Creek	52.4219	-126.1586	3,720							
08FC003	Dean River below Tanswanket Creek	52.8897	-125.7714	3,780							
08GD004	Homathko River at the Mouth	50.9847	-124.9169	5,720	**18**	South Pacific	08FA002 + 08FB006 + 08FB007 + 08FC003 + 08GD004	51.9497	-125.4700	19,550	1,432
Fraser-Okanagan-Columbia Rivers											
08MF005	Fraser River at Hope	49.3806	-121.4514	217,000	**19**	Fraser	08MF005	52.5624	-122.4017	217,000	1,107
08NM085 [R]	Okanagan River near Oliver	49.1147	-119.5639	7,590	**20**	Okanagan	08NM085	49.8509	-119.5030	7,590	1,185
08NE058 [R]	Columbia River at International Boundary	49.0008	-117.6278	155,000	**21**	Columbia	08NE058	50.2890	-116.7548	155,000	1,590

Table 3.2(con't). Details of the CROCWR study region.

WSC Station					***CROCWR Watersheds***					
Station ID	**Station Name**	**Latitude**	**Longitude**	**Drainage Area (km^2)**	**Watershed Name**	**Definition (based on WSC stations)[%]**	**Latitude***	**Longitude***	**Drainage Area (km^2)**	**Median Basin Elevation (m.a.s.l.)**
Saskatchewan River										
05DF001 [R]	North Saskatchewan River at Edmonton	53.5369	-113.4853	28,100	**22** Upper North Saskatchewan	05DF001	52.5500	-115.7790	28,100	1,337
05GG001 [R]	North Saskatchewan River at Prince Albert	53.2028	-105.7683	131,000	**23** Lower North Saskatchewan	05GG001 - 05DF001	52.8366	-110.2368	102,900	669
05AJ001 [R]	South Saskatchewan River at Medicine Hat	50.0419	-110.6775	56,400						
05CK004 [R]	Red Deer River at Blindloss	50.9028	-110.2972	47,850	**24** Upper South Saskatchewan	05AJ001 + 05CK004	50.7934	-113.3094	104,250	957
05HG001 [R]	South Saskatchwan River at Saskatoon	52.1406	-106.6442	141,000						
05KJ001 [R]	Saskatchewan River at the Pas	53.8417	-101.1861	389,000	**25** Lower South Saskatchewan	05KJ001 - 05GG001 - 05HG001	53.6606	-103.9974	117,000	398

[N] Station is part of the RHBN
[R] Station is located on a regulated river
[+] Station data are amalgamated with station 07LE001 data
[++] Station data are amalgamated with station 07JD001 data
[+++] Station data are amalgamated with station 08BB001 data
[#] Station data are formed through combination of stations 10GC001, 10ED002, 10FB002, and 10FA002, accounting for travel time
[%] Watershed streamflow data are derived from defined formula of output station(s) data minus input station(s) data (where some watersheds have no input station(s))
* Coordinates taken at centroid of each watershed

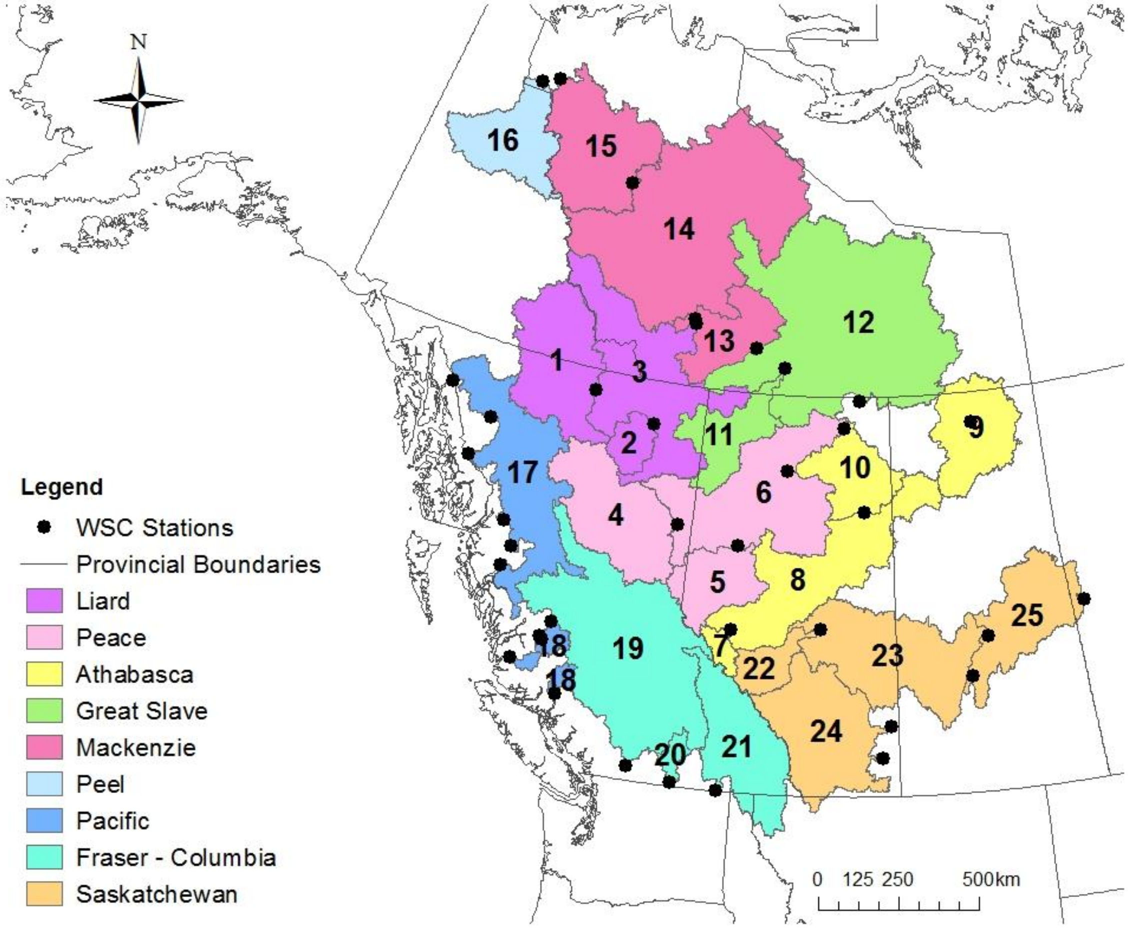


Figure 3.3. Map of the CROCWR study region by watershed. Refer to Table 3.2 for corresponding names and properties of each watershed.

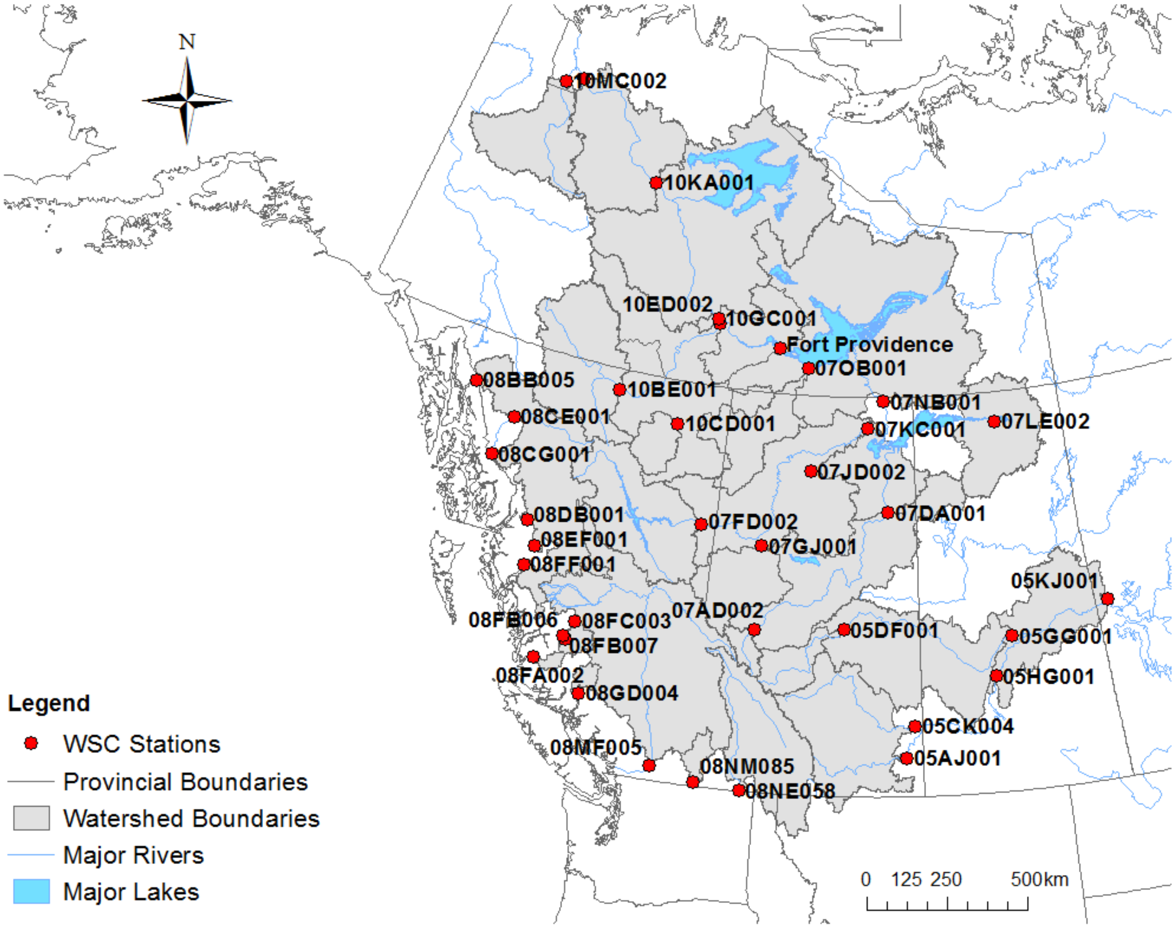


Figure 3.4. CROCWR study region by station. Major rivers and lakes are marked. Refer to Table 3.2 for corresponding properties of each station.

Chapter 4

Data and Methods

This research involved analysis of both hydrological and climatic data. This chapter outlines the data and research methods that were applied in the study. *Section 4.1* introduces the hydrological and climatic variables analyzed and the various data sources from which they were obtained. A description of how each of the variables was defined is included. *Section 4.2* describes the methodologies used in this research. This includes information regarding the Mann-Kendall test for trend, hydrological regionalization by PCA, and determining hydro-climate linkages through correlation and composite analyses.

4.1 Data and Sources

4.1.1 Streamflow Data

4.1.1.1 Runoff Variables

Streamflow can be used to estimate the amount of runoff produced within a watershed. Daily streamflow data for each WSC station were acquired via the WSC Archived Hydrometric Database (HYDAT) (Environment Canada, 2013a) for two different periods: 1976-2010 (35 years) and 1966-2010 (45 years). Due to limitations in the lengths of the WSC hydrometric records for the majority of the CROCWR stations, especially in the North, it was decidedly not useful to perform analyses for longer periods of time. The two periods selected reflect a trade-off between greater spatial coverage with the shorter period, since more stations have full data records, versus greater power for the statistical tests for the longer period, despite many stations exhibiting data gaps. The shortest time period also represents conditions strictly following the 1976/1977 PDO regime shift, a significant period at which a number of hydrological and ecological changes in western North America were documented due to a shift in climate from cool, wet conditions to warm, dry conditions (Mantua and Hare, 2002; Gobena and Gan, 2006); the longer analysis period captures more persistent hydrological trends, both preceding and encompassing the mid-1970s shift.

Daily streamflow series for each CROCWR watershed were produced based on the definition of station input(s) and output(s) provided in Table 3.2. Annual and seasonal discharge series were created by averaging the daily series. Seasons were defined as winter (January, February, March), spring (April, May, June), summer (July, August, September), and fall (October, November, December) based on a review of the watershed hydrographs. Although this definition differs slightly from the more commonly used definition of winter (December, January, February), spring (March, April, May), summer (June, July, August), and fall (September, October, November), March flows were lower than January flows in 17/25 of the basins, therefore it was deemed appropriate to shift the seasons such that the lowest flows occurred during winter months for the majority of the study area. Six-month cold and warm season variables were also created; the cold season was defined as the months of November to April, while the warm season was specified as the months of May to October, inclusive. Annual flow was calculated based on the hydrological year (October to September). Annual and seasonal discharge data for each watershed were then divided by their respective drainage areas to produce runoff time series in units of mm. This was done to remove the effects of watershed size on calculated statistics (Déry *et al.*, 2009). Figure 4.1 shows the effect of this transformation. Note that inter-watershed variability is reduced by dividing by watershed area.

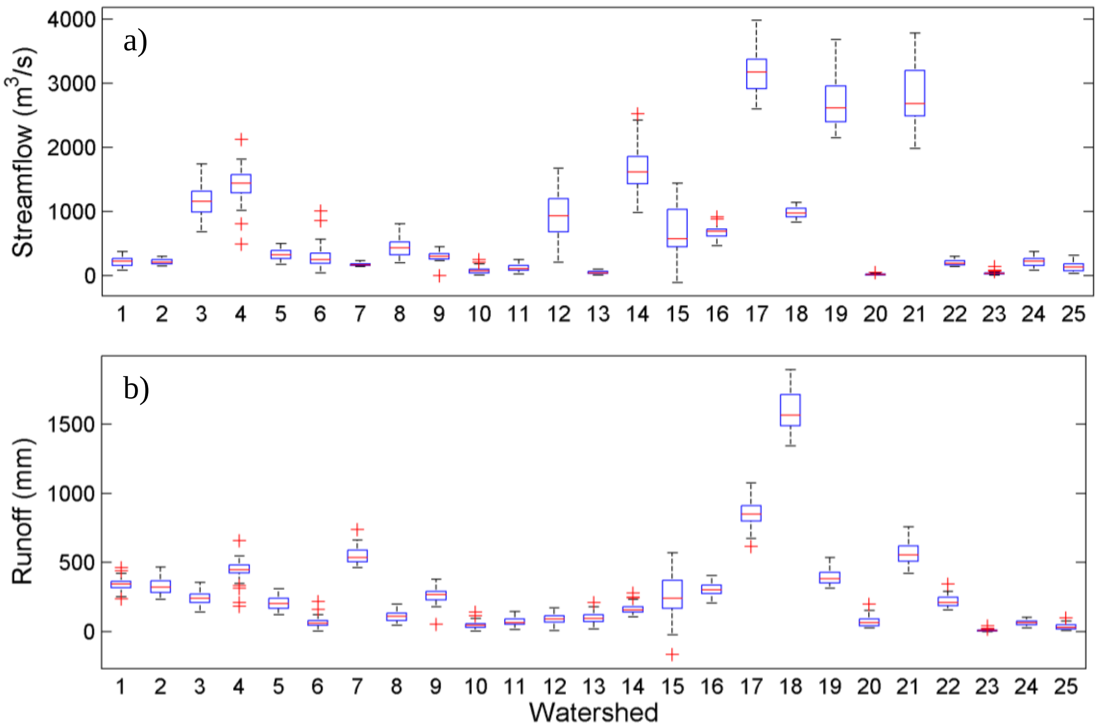


Figure 4.1. Box plots of a) mean annual streamflow and b) mean annual runoff for each of the CROCWR watersheds for the 1976-2010 time period.

Table 4.1 presents a summary of mean annual streamflow and mean annual runoff from 1976-2010 for each of the CROCWR watersheds. Appendix A, Table A.1 provides this information for the period of 1966-2010.

4.1.1.2 Streamflow Timing Variables

In addition to runoff, it was of interest to explore changes in the timing of significant hydrological events in the CROCWR region. The beginning and end dates of the spring freshet were calculated using the daily streamflow records. The two periods of 1976-2010 and 1966-2010 were again used for streamflow timing analysis.

Table 4.1. Mean annual streamflow and mean annual runoff values for the CROCWR study region for the period of 1976-2010.

Watershed		Annual Streamflow (m^3/s)	Annual Runoff (mm/yr)
1	Upper Liard	1138.2	346.1
2	Fort Nelson	216.4	337.5
3	Lower Liard	1148.2	241.3
4	Upper Peace	1458.2	455.7
5	Smoky River	316.5	199.0
6	Lower Peace	301.3	68.3
7	Upper Athabasca	168.0	544.6
8	Lower Athabasca	421.4	108.6
9	East Lake Athabasca	303.3	268.5
10	West Lake Athabasca	74.4	47.8
11	Hay	123.0	75.2
12	Great Slave	967.4	94.8
13	Upper Mackenzie	48.0	100.8
14	Mid Mackenzie	1559.5	154.2
15	Lower Mackenzie	638.4	228.6
16	Peel	676.9	305.3
17	North Pacific	3177.6	851.7
18	South Pacific	982.7	1589.6
19	Fraser	2663.5	388.9
20	Okanagan	18.4	76.6
21	Columbia	2749.2	560.3
22	Upper North Saskatchewan	193.6	217.4
23	Lower North Saskatchewan	31.1	9.5
24	Upper South Saskatchewan	202.8	61.3
25	Lower South Saskatchewan	132.1	35.3

The start of the spring freshet has been defined in a number of different ways. In an analysis of streamflow timing in the headwaters of the MRB, Burn (2008) outlined four different measures of the start of the spring freshet: the five percentile (5P) date, the ten percentile (10P) date, the “spring freshet” date, and the spring pulse date. The 5P and 10P dates are defined as the Julian dates of the calendar year by which 5% and 10% of the annual streamflow volume has occurred, respectively. The “spring freshet” date is derived through a combination of automated and manual analyses, and involves determining the first date on which streamflow exceeds 1.5 times the average of the preceding 16 days and revising the dates which do not represent the start of the spring freshet by visual examination of the hydrograph. The spring pulse onset date (Cayan *et al.*, 2001) is defined as the Julian date of the calendar year on which the cumulative departure of streamflow from the annual average is the most negative, and is equivalent to determining the date after which the majority of streamflows are above the yearly average. Figure 4.2 provides a plot of how the spring pulse date was determined for one station and one year. Burn (2008) determined that the "spring freshet" and 10P dates were more effective measures of the start of the spring freshet than were the 5P or pulse dates, although the pulse date has been widely used in other research with convincing results (Cayan *et al.*, 2001; Stewart *et al.*, 2004). Due to the ability to fully automate the calculations, the 10P and spring pulse dates were calculated in this work.

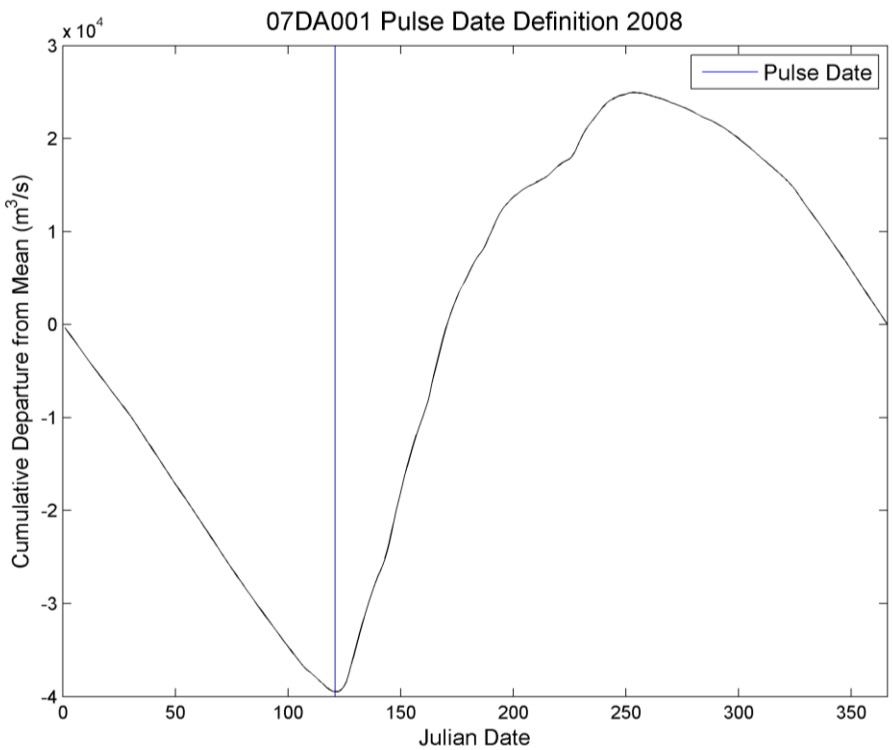


Figure 4.2. Cumulative departures from the mean for daily flows at Athabasca River below McMurray (WSC station 07DA001) for the year 2008. The spring pulse date is defined on the day when the cumulative departure from the mean is minimum, and is shown by the blue line. The pulse date in 2008 was Julian day 121 (April 30).

The end of the spring freshet can be gauged by determining the hydrograph centre of mass (CM) (Stewart *et al.*, 2004). The timing of the CM of annual flow is calculated as:

$$CM = \frac{\sum(t_i q_i)}{\sum q_i} \tag{4.1}$$

where t_i is the time in days from the beginning of the calendar year and q_i is the corresponding streamflow for day i. Johnson (1964) argued that April 1 (rather than January 1) should be used as a start date in the calculation of the CM to remove the effects of winter variations due to rainfall or unusual, early melt; Whitfield (2013a) recommended that streamflow separation be conducted to isolate the snowmelt component of the hydrograph so that the CM does not include the effects of other processes. Prowse (2013, personal communication) suggested using the onset of the spring freshet as the start date in the calculation of the CM so as to ensure that winter low flows are excluded from the calculation, which could otherwise lead to an earlier calculated CM and hence a seemingly shorter spring freshet. The method of Prowse (2013, personal communication) was used to calculate the CM as a measure of the date of the end of the spring freshet.

Figure 4.3 shows the 10P date, pulse date, and two definitions of the CM date on a daily flow hydrograph for one station and one year. The length of the spring freshet was calculated by subtracting the date of the start of the spring freshet from the CM date. The volume of the spring freshet was defined as the runoff (in mm) that occurred between the start and end dates of the freshet; this again ensured that watershed area would not lead to incomparable results among watersheds of different sizes.

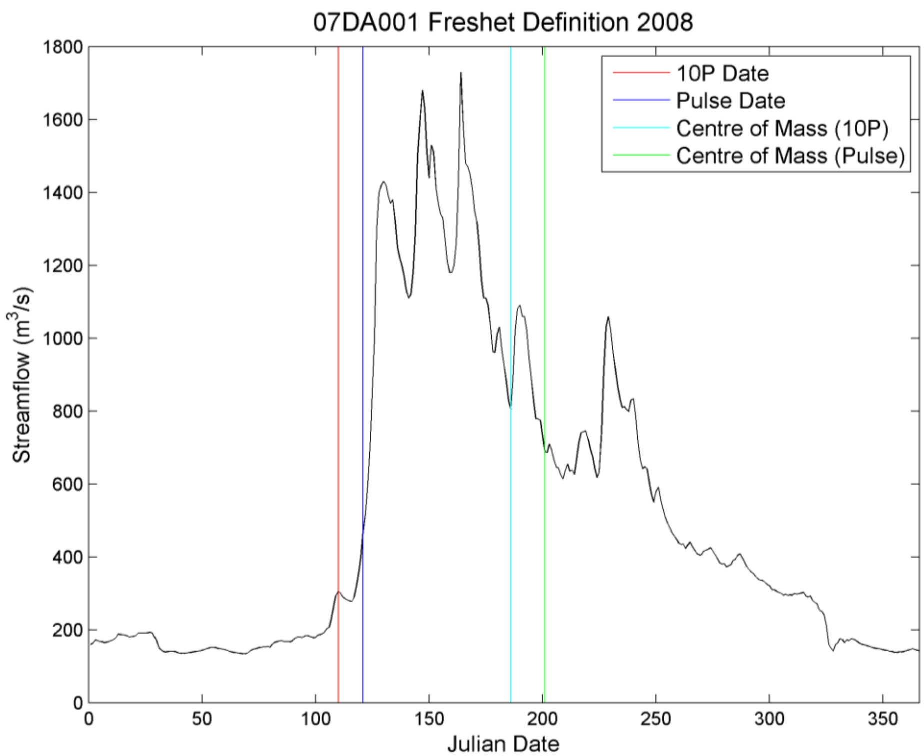


Figure 4.3. Daily flows at Athabasca River below McMurray (WSC station 07DA001) for the year 2008. The 10P date is shown by the red line (Julian day 110 or April 19); the spring pulse date is shown by the blue line (Julian day 121 or April 30); and the CM date is shown by the cyan line using the 10P date as starting date (Julian day 186 or July 4) and by the green line using the pulse date as starting date (Julian day 201 or July 19).

4.1.2 Surface Climate Data

As discussed in *Chapter 1*, the local water budget of a watershed is controlled by both precipitation and air temperature. Precipitation provides direct input to streamflow and runoff volume as liquid (rain) or frozen (sleet, hail, or snow) water. The influence of air temperature, however, is less obvious. Air temperature controls the water cycle as an index of evaporation. Evaporation includes all processes by which water returns to the atmosphere as water vapour (Winkler *et al.*, 2010). The rate of sublimation, or the "evaporation" of snow without first passing through the liquid phase, is also controlled by air temperature. Consequently, the volume of water stored as snowpack or glacial ice, the primary sources of freshwater in western Canada, are intricately linked to air temperature. Changes in air temperature can also lead to changes in the timing of melt of these frozen water stores. It is therefore important to consider surface climate when analyzing trends in streamflow/runoff.

A number of different meteorological variables were analyzed to investigate the links between hydrology and surface climate. The ANUSPLIN interpolated dataset developed by McKenney *et al.* (2011) and introduced in *Section 2.2.1* was acquired from Natural Resources Canada for this purpose. This data set contains daily records of maximum temperature, minimum temperature, and total precipitation for every (10 km × 10 km) grid point in Canada. ArcGIS was used to extract and spatially average the data over each CROCWR watershed, thus creating daily time series of temperature (maximum and minimum) and precipitation for each basin. In addition, a mean temperature data set was created for each basin by averaging maximum and minimum values at each point and each time interval. Similar to streamflow data, daily temperature and precipitation data were averaged into monthly, seasonal, and annual time series. Annual maximum, mean, and minimum daily temperatures and total precipitation averaged over the 1976-2010 period for each watershed are given in Table 4.2.

Table 4.2. Average maximum, mean, and minimum daily temperatures and total annual precipitation values for the CROCWR study region for the period of 1976-2010.

	Watershed	Max. Temp. (°C)	Mean Temp. (°C)	Min. Temp. (°C)	Total Annual Precip. (mm)
1	Upper Liard	18.5	-3.2	-27.7	509.9
2	Fort Nelson	19.0	-1.3	-24.1	562.4
3	Lower Liard	20.4	-2.8	-28.2	467.9
4	Upper Peace	18.8	0.3	-20.3	586.5
5	Smoky River	20.8	2.0	-20.1	538.0
6	Lower Peace	21.9	0.3	-25.0	425.0
7	Upper Athabasca	16.6	-0.5	-18.8	711.6
8	Lower Athabasca	22.2	1.4	-22.6	474.6
9	East Lake Athabasca	21.3	-3.7	-30.4	436.7
10	West Lake Athabasca	22.8	-0.3	-26.4	399.9
11	Hay	22.0	-1.2	-27.7	408.8
12	Great Slave	20.2	-5.1	-31.3	295.7
13	Upper Mackenzie	22.3	-3.5	-30.3	326.0
14	Mid Mackenzie	19.2	-6.7	-31.9	294.8
15	Lower Mackenzie	19.8	-7.0	-32.5	286.5
16	Peel	18.7	-6.6	-30.8	353.2
17	North Pacific	17.2	0.2	-18.1	807.9
18	South Pacific	17.9	1.3	-15.2	766.2
19	Fraser	20.9	2.5	-15.6	628.1
20	Okanagan	24.2	4.8	-11.2	550.5
21	Columbia	20.2	1.6	-14.5	918.3
22	Upper North Saskatchewan	19.8	1.4	-18.9	611.5
23	Lower North Saskatchewan	24.7	2.5	-21.9	392.2
24	Upper South Saskatchewan	24.2	3.7	-17.9	446.6
25	Lower South Saskatchewan	24.4	0.9	-25.3	425.3

4.1.3 Atmospheric Climate Data

Connections between runoff and various climate indices (teleconnections) were also explored to determine if linkages exist. A climate index is a diagnostic quantity that is used to characterize an aspect of a geophysical system, such as a circulation pattern (NCAR, 2012). Certain indices exhibit more prominent influences on climate in different regions of the world. The primary Pacific teleconnections affecting Canadian climate include the El Niño-Southern Oscillation

(ENSO), the Pacific Decadal Oscillation (PDO), the Pacific North American (PNA) pattern, and the North Pacific (NP) index (Bonsal *et al.*, 2006). The Arctic Oscillation (AO) has been reported to have strong influence on conditions over the central Arctic Ocean, as well as with rivers draining into Hudson Bay (Déry and Wood, 2004), and hence was also included in this analysis. The Aleutian Low, an integral part of these Pacific-based teleconnections, is known to control the degree that heat and moisture is advected over western Canada. The Aleutian low is a semi-permanent low pressure center located near the Aleutian Islands during the winter. A deepening of the Aleutian low over Alaska is usually accompanied by higher-than-normal pressure over western Canada; as the low moves east, warm and moist air generally occurs over northwestern Canada (e.g., Hartmann and Wendley, 2005), and is often associated with cyclone formation or intensification (e.g., Spence and Rausch, 2005).

The ENSO is one of the best known and most prominent climate signals for which teleconnections to land surface variables affecting hydrology have been documented (Maurer *et al.*, 2004; Burn, 2008). The ENSO is an important source of inter-annual variability in weather and climate with a periodicity of 6 to 18 months (Manthua and Hare, 2002) and includes both warm (El Niño) and cool (La Niña) phases. The SOI index from the US National Oceanographic and Atmospheric Administration (NOAA) Climate Prediction Centre was used to characterize ENSO in this study (http://www.esrl.noaa.gov/psd/data/correlation/soi.data). The SOI index is based on differences in standardized sea level pressure (SLP) values between Tahiti, French Polynesia and Darwin, Australia. Winter season SOI is an indicator of the mature phase of the ENSO (Gobena and Gan, 2006).

The PDO (Mantua *et al.*, 1997) is a pattern of North Pacific climate variability that shifts phases on inter-decadal time scales, usually about every 20 to 30 years. The PDO index is defined as the leading PC of monthly sea surface temperature (SST) anomalies in the North Pacific Ocean poleward of 20°N. Similar to the effects of ENSO, western North American climate is influenced by the different phases of PDO, with warm, dry conditions associated with a positive PDO and cool, wet conditions resulting during a negative PDO. The Aleutian low tends to be strong during the positive phase of the PDO (Serreze *et al.*, 2008). The most well-documented shift in the PDO regime occurred between 1976 and 1977, when a cool period ended and a warm period began, although a shift back to a cool phase also occurred around 1998 (Mantua and Hare, 2002;

Gobena and Gan, 2006) (or 2001, as per Hartmann and Wendler, 2005). The PDO index was acquired from NOAA (http://www.esrl.noaa.gov/ psd/data/correlation/pdo.data).

The PNA pattern describes mid-tropospheric circulation over the North Pacific and North America (Bonsal *et al.*, 2006). The PNA is defined as a measure of atmospheric response to a warm SST anomaly in the central equatorial Pacific. The PNA has been found to be a dominant mode of variation in the North Hemisphere mid-latitudes during the winter months, and is strongly associated with North American temperature and precipitation anomalies. Similar to PDO, strongly positive (negative) PNA indices are associated with El Niño warm (La Niña cold) events (Coulibaly and Burn, 2005). The PNA index was acquired from NOAA (http://www.esrl.noaa.gov/psd/data/correlation/pna.data).

The NP index provides a measure of the strength of the Aleutian low during the winter period, and was found to relate to streamflow variability in the western United States (Lins, 1997). The NP can also be considered a SLP-based index depicting the PNA pattern, and is essentially a mirror image of the PNA (Maurer *et al.*, 2004). The NP index was acquired from the NCAR Climate Data Guide (http://climatedataguide.ucar.edu/sites/default/files/cas_data_files/asphilli/npindex_monthly_1.txt).

The AO is defined as the leading PC of SLP variability north of 20° latitude. Negative AO events are indicative of higher SLP over the polar region and lower SLP over the mid-latitudes, and are often associated with a southward shift in the jet stream allowing cold air surges into the mid-latitudes. Positive AO events include a northward shift in the jet stream and fewer cold air surges into the mid-latitudes (Bonsal *et al.*, 2006). The AO index was acquired from NOAA (http://www.esrl.noaa.gov/psd/data/correlation/ao.data).

4.2 Research Methods

4.2.1 Trend Detection

4.2.1.1 Change-Point Detection

Prior to testing for trend, the data were assessed for change-points in the time series. The Pettitt non-parametric change-point test (Pettitt, 1979) was used to solve the change-point problem. In a sequence of random variables X_1, X_2, …, X_T, the sequence is said to have a change-point at time τ if X_t for t = 1,…,τ have a common distribution function $F_1(x)$ and X_t for t = τ + 1,…,T have a common distribution function $F_2(x)$, where $F_1(x) \neq F_2(x)$. Similar to the MK test, the null hypothesis H_0 assumes that the data are homogeneous (IID) such that there is no change in the distribution of the data at any time t. The alternative hypothesis, H_1, states that there is a break in the mean of the data series at time τ, and therefore the data are inhomogeneous. No assumption is made about the functional forms of F_1 and F_2 except that they are continuous.

Theoretically, a change point represents a sudden change in the statistics of a record. If the two samples $(x_1, x_2, \dots, x_t)$ and $(x_{t+1}, x_{t+2}, \dots, x_T)$ come from the same population, the test statistic $U_{t,T}$ is given by:

$$U_{\tau,T} = \sum_{i=1}^{t} \sum_{j=t+1}^{T} sgn(x_t - x_j) \tag{4.2}$$

The most significant change-point is found where the value of $|U_{t,T}|$ is max: $K_T = \max|U_{t,T}|$. The significance level associated with K_T^+ or K_T^- is approximately determined by (Zhang *et al.*, 2009):

$$\rho = exp\left(\frac{-6K_T^2}{T^3 + T^2}\right) \tag{4.3}$$

If ρ is smaller than the specific significance level, e.g. 0.10 (or 10%) in this study, the null hypothesis is rejected. In other words, if a significant change point exists, the time series is divided into two parts at the location of the change point (Zhang *et al.*, 2009).

4.2.1.2 Mann-Kendall Non-Parametric Trend Test

As alluded to in *Section 2.2.2*, the MK test is a common method used in hydroclimatic research for detecting trend in a time series, and was therefore applied in this work. The MK test is a rank-based method and hence is robust to non-normality of the underlying data. According to Mann, the null hypothesis H_0 states that the data are a sample of N independent and identically distributed (IID) random variables – that is, that there is no trend present in the data. The alternative hypothesis H_1 is that the distribution of x_i and x_j is not identical for all $i, j \leq N$ with $i \neq j$, where x_i and x_j are values of the time series. In other words, H_1 assumes that a monotonic trend is present in the time series. In a two-sided test for trend at significance level α, H_0 should be rejected if $|Z| \geq z_{\alpha/2}$, where $F_{N}(z_{\alpha/2})$ is the standard normal cumulative distribution function and Z is the test statistic:

$$Z = \begin{cases} \dfrac{S-1}{\sqrt{Var(S)}} & if\ S > 0 \\ 0 & if\ S = 0 \\ \dfrac{S-1}{\sqrt{Var(S)}} & if\ S < 0 \end{cases} \tag{4.4}$$

where S is defined by:

$$S = \sum_{j=1}^{N-1} \sum_{i=j+1}^{N} sgn(x_i - x_j) \tag{4.5}$$

and:

$$sgn(\theta) = \begin{cases} 1\ if\ \ \theta > 0 \\ 0\ if\ \ \theta = 0 \\ -1\ if\ \ \theta < 0 \end{cases} \tag{4.6}$$

The test statistic S is a count of the number of times x_i exceeds x_j for $i > j$ more than x_j exceeds x_i. The maximum possible value of S occurs when $x_1 < x_2 < \ldots < x_n$ (Hipel, 2011).

Kendall obtained the theoretical mean and variance of S under the assumption of no trend as:

$$E(S) = 0 \tag{4.7}$$

$$Var(S) = \frac{[n(n-1)(2n+5) - \sum_t t(t-1)(2t+5)]}{18} \tag{4.8}$$

where t is the extent of any tie. A continuity correction for improving the approximation to normality is recommended when less than 10 data values are available.

Trend slopes were computed using a robust non-parametric slope estimator developed by Theil (1950) and Sen (1968), which hereinafter shall be referred to as the Theil-Sen approach (TSA):

$$\beta = median\left\{\frac{x_i - x_j}{i - j}\right\} \quad \text{for all } j < i \tag{4.9}$$

where β is the slope estimator and x_i and x_j are flows at time i and j, respectively.

4.2.1.3 Trend-Free Pre-Whitening

The data used in this analysis were corrected for serial correlation by means of the Trend Free Pre-Whitening (TFPW) approach developed by Yue *et al.* (2002). The TFPW involves estimating a monotonic trend for the series, removing this trend prior to pre-whitening the series, and finally adding the monotonic trend calculated in the first step to the pre-whitened data series. In essence, the TFPW approach attempts to separate the serial correlation that arises from a (linear) trend from the remaining serial correlation and then only removes the latter portion of the serial correlation. Although the TFPW procedure involves fitting and removing a linear trend, the MK trend detection procedure does not make any assumptions about the nature of the trend in a data set. The MK test was performed at both the 5% and 10% significance levels.

Below are the steps involved in removing serial correlation (Burn *et al.*, 2004a):

Step 1: Estimate the MK test statistic S and evaluate the local significance level α of the trend in the original data series x_t. Calculate the non-parametric slope β of the data using the TSA. If β is almost equal to zero, then it is not necessary to conduct trend analysis. If β differs from zero, remove the monotonic (linear) trend from the data series through:

$$y_t = x_t - \beta t \tag{4.10}$$

where x_t is the series value at time *t* and y_t is the de-trended series.

Step 2: Evaluate r_1, the lag-one serial correlation coefficient of the de-trended series y_t. If the value of r_1 is not statistically significant at significance level α, the trend results from Step 1 are used and the calculations for the data set are complete. If the serial correlation is significant, pre-whiten the de-trended series through:

$$y_t' = y_t - r_1 y_{t-1} \tag{4.11}$$

where y_t' is the de-trended and pre-whitened series, and is referred to as the residual series.

Step 3: Add the monotonic trend to the residual series through:

$$y_t'' = y_t' + \beta t \tag{4.12}$$

where y_t'' is the trend-free pre-whitened series.

Step 4: Calculate the MK test statistic S' and local significance of the calculated S' for the blended series y_t''.

4.2.1.4 Field Significance

The field significance of each variable was evaluated using Walker's test (Wilks, 2006). If there are *K* sites for which the (local) significance has been evaluated, Walker's test involves determining the smallest of the *K* p-values that have been determined and comparing this value to a critical value, which depends on the (global) significance level selected and the number of tests, *K*. Wilks (2006) demonstrated the relationship between Walker's test and the concept of the false discovery rate, where the latter considers all *K* local p-values. Walker's test considers the magnitude of the p-values in determining field significance and hence the strength of the trend at each individual station is explicitly considered. Since this is a two-sided test, it does not distinguish between the tails of probability curve, i.e. whether one collection of trends (increasing or decreasing) is field significant and the other is not. Therefore, if a variable is found to be field significant, then the entire set of significant trends (increasing and decreasing) is field significant. Walker's test for field significant was evaluated at the 5% and 10% significance levels.

4.2.2 Hydrological Regionalization by PCA

PCA was used to distinguish independent regions of coherent hydrological variability in the CROCWR study region. In PCA, a set of intercorrelated variables is transformed into a new set of mutually orthogonal eigenvectors, or modes, called principal components (PCs); PCs are linear combinations of the original variables. PCA attempts to place a large portion of the total variance in the first PC (dominant mode), and each successive PC claims a smaller portion of the variance. The time series related to the first few eigenvectors represent a large fraction of variability of the original dataset.

In PCA, the dominant modes of spatial variability are expressed as the PCs of either a correlation or covariance matrix. The correlation matrix standardises between-variable variance such that it has zero mean and unit variance, whereas the covariance matrix retains non-standardised variance (Li, 2012). There is no universal rule on deciding whether a correlation or a covariance dispersion matrix is preferable, although Lins (1997) and Maurer *et al.* (2004) recommend the use of the correlation matrix in hydrological research to eliminate any tendency for skewed or overweighted spatial patterns resulting from point or regional differences in variance magnitude.

The eigenvectors in a PCA illustrate how geographical areas of runoff vary together. In some cases, these spatial patterns become more easily interpretable and physically meaningful when the eigenvectors are subjected to a rotation. There are two general types of transformations: orthogonal and oblique rotations. Orthogonally rotated solutions ensure that PCs remain uncorrelated, whereas oblique rotations relax this requirement to achieve greater interpretability. Both rotations have been widely used in the field of hydrometeorology. White *et al.* (1991) found that oblique rotations were more stable than both orthogonally rotated and unrotated solutions, whereas Miller and Goodrich (2007) found little difference between the two types and preferred the orthogonal method because it tends to preserve maximum loading of individual components (Chen *et al.*, 2009). The orthogonal Varimax rotation is commonly used in PCA on hydroclimatic data (Lins, 1997; Maurer *et al.*, 2004; Romolo *et al.*, 2006; Jonsdottir and Uvo, 2009). Orthogonal rotation of the eigenvectors results in redistribution of the variances among the rotated modes (Jonsdottir and Uvo, 2009).

Choosing the right number of PCs to retain can present a challenge as the problem does not have one single, rigorous solution. Many techniques have been developed to evaluate the 'optimal' number of components, including both heuristic and statistical methods. Morrison (2005) recommended that the total variance accounted for by the chosen components be at least 80%, but that the number of components be no more than five. The Kaiser-Guttman criterion suggests retaining all components with eigenvalue greater than magnitude one. The scree plot is a common visual method used for choosing which PCs to retain, and involves plotting the value of each successive eigenvalue against the rank order and noting where the descent of the plot slows abruptly. Interpretable components locate to the left of the plot 'elbow', whereas smaller eigenvalues, representing random variation, tend to lie along a straight line to the right of this point and can be dropped from the analysis (Jackson, 1993). Other methods for determining how many PCs to keep include the broken-stick model, Bartlett's test of sphericity, Bartlett's test of homogeneity of the correlation matrix, Lawley's test of the second eigenvalue, bootstrapped confidence limits on successive eigenvalues, and bootstrapped confidence limits on eigenvector coefficients (Jackson, 1993).

When dealing with space-time data series, there are six different operational modes of PCA: O, P, Q, R, S, and T (Demsar *et al.*, 2013). The S-mode decomposition concerns one location varying over time and is the mode most commonly used in the atmospheric sciences and hydrology. An S-mode PCA is typically performed on an $n \times p$ data matrix to isolate subgroups of locations (e.g. hydrometric stations) that tend to co-vary similarly, where p is the number of locations and n is the number of time-observations (Piechota *et al.*, 1997).

An S-mode PCA was used to classify distinct areas of runoff and streamflow timing variability within the CROCWR region. The correlation matrix was specified in the PCA to avoid loadings being focused exclusively in regions of high runoff (Maurer *et al.*, 2004); the results therefore represent patterns of spatially coherent runoff (or timing) anomalies (Lins, 1997). Accordingly, the data matrix was structured with time (in years) as rows and watersheds (or stations, in the case of spring freshet) as columns.

Regarding the decision of whether to perform rotation of the PCA, it is believed that a rotated solution is superior to a non-rotated solution because rotation facilitates physical interpretation of the PCs (Yarnal, 1993), and is deemed necessary when PCA is applied in a spatial context (Richman, 1986). The PCA solutions were therefore subjected to the Varimax orthogonal rotation. Inspection of pre- and post-rotated PCA results verified this recommendation.

The scree plot was used to determine the number of PCs to retain. To facilitate comparison among the solutions for annual, monthly, and seasonal runoff (or timing), the same number of components were rotated and analyzed for each of the variables. Figure 4.4 shows the pre- and post-rotation scree plots for 1976-2010 annual runoff PCA. On each graph, left-side y-axis is eigenvalue, while right-side y-axis is cumulative percent variance explained.

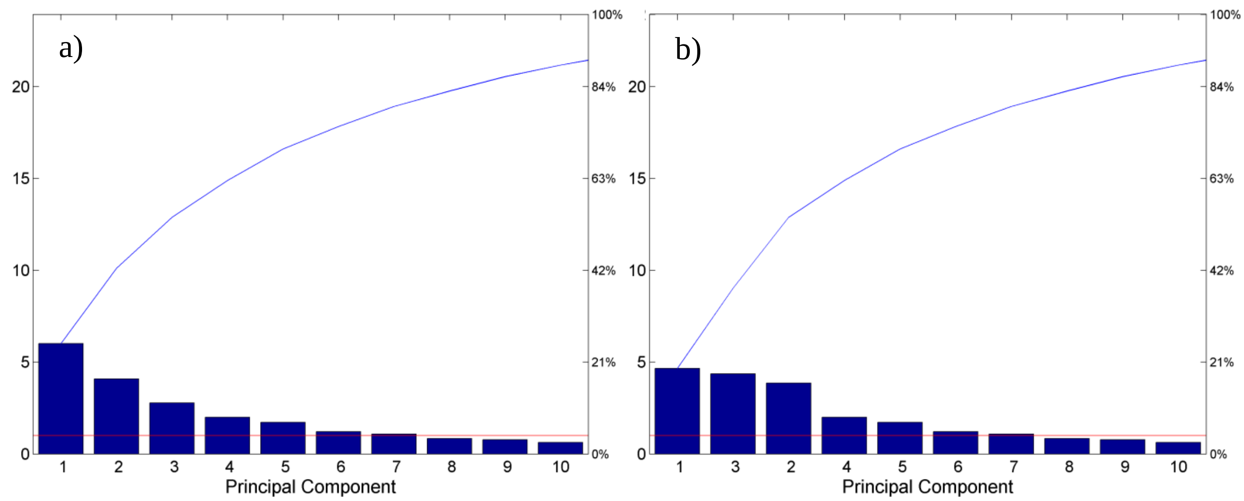


Figure 4.4. Scree plots for a) pre-rotated PCA solution and b) three-component Varimax orthogonally rotated PCA solution for 1976-2010 annual runoff data.

4.2.3 Correlation Analysis

Direct linear correlation analyses were performed between standardized runoff series and standardized maximum temperature, minimum temperature, mean temperature, and total precipitation series for each of the CROCWR watersheds and for each variable (i.e. annual, monthly, seasonal). For the timing of the start and end of the spring freshet, correlation analyses were performed between standardized hydrological timing variables at each hydrometric gauging station and the standardized temperature and precipitation series of the respective contributing (upstream) watershed.

The Pearson correlation coefficient is calculated by:

$$r = \frac{\sum_i (x_i - \bar{x})(y_i - \bar{y})}{\sqrt{\sum_i (x_i - \bar{x})^2}\sqrt{\sum_i (y_i - \bar{y})^2}} \tag{4.13}$$

where *x* and *y* are two data sets (arrays) of the same length (e.g. *x* is runoff, *y* is precipitation) and $\bar{x}$ and $\bar{y}$ are the means of the respective series.

Temperature and precipitation do not always produce a direct or immediate response on runoff or streamflow timing. For this reason, the correlation coefficient was assessed using a number of different lead times. Lead time indicates the number of intervening seasons (or months) between the season (or month) during which the temperature or precipitation occurred, and the season (or month) when the runoff (or hydrological event) occurred. By including different lead times in the analysis, one can gauge how far in the past changes in climate can affect changes in runoff or timing, and thus predict future changes in runoff or timing knowing today's climate. For example, April runoff may be correlated with March precipitation and February temperature.

4.2.4 Composite Analysis

The "circulation-to-environment" composite analysis approach was used to link runoff to atmospheric circulation patterns. The SOI (representing the ENSO), PDO, PNA, NP, and AO climate indices were analyzed. Following the method of Maurer *et al.* (2004) and Burn (2008), the 8 largest and 8 smallest values for each climate index were selected and standardized runoff values (or timing dates) for each watershed (or station) were examined for the years associated with each of the selected climate index values by means of a t-test. Similar to the correlation analysis method, relationships were tested at a series of lead times, since atmospheric patterns are generally known to exert greater influence on surface climate and hydrology during certain times/seasons than others. For example, typical ENSO events tend to develop during summer to early fall, mature during winter, and terminate the following spring, but have been shown to be associated with a higher number of extended summer dry spells in the Canadian Prairies during the second summer following the mature stage (Bonsal and Lawford, 1999).

Chapter 5

Results

This chapter presents the results of exploratory statistical analysis relating to western Canadian water resources. *Section 5.1* outlines the results of MK trend analysis for both runoff and streamflow timing data, including analysis of regime shifts and characterization of the spring freshet variables. *Section 5.2* details the results of PCA for each variable, including an explanation of regions exhibiting similar hydrological behaviour and trend analysis of the projected time series for each region. *Section 5.3* summarizes the results of climatic analysis, including trend results of the meteorological variables studied, correlation analysis between surface climate variables and streamflow data, and composite t-test analysis with climatic indices.

5.1 Hydrological Trend Results

MK trend analysis was applied to both runoff and streamflow timing series for two periods of time: 1976-2010 (35 years) and 1966-2010 (45 years). As discussed in *Section 4.1.1.1,* two analysis periods were used in order to examine shorter-term trends over the broader study region, as well as longer-term trends in watersheds/at stations where longer-term records were available.

Figure 5.1 shows two examples of basins whose 1976-2010 and 1966-2010 average hydrographs differed notably, where peak flows tended to be lower during the later/shorter period.

For the 45-year analysis period, a data set was only considered for analysis if there were no more than four years of missing data within the analysis period (e.g., Burn and Hag Elnur, 2002). This condition was relaxed for the 1976-2010 period in order to be able to present results for all basins/stations for at least the shortest time period. Missing data were not in-filled based on the recommendation of Whitfield (2013b). As a result, the nominal number of basins/stations for the 45-year period was less than the total number of basins/stations overall due to missing data for some variables for some basins/stations.

The zyp package for R Statistical Software (http://cran.r-project.org/web/packages/zyp/index.html) was used to test for trend in the data; the Yue-Pilon TFPW method was specified in the analysis.

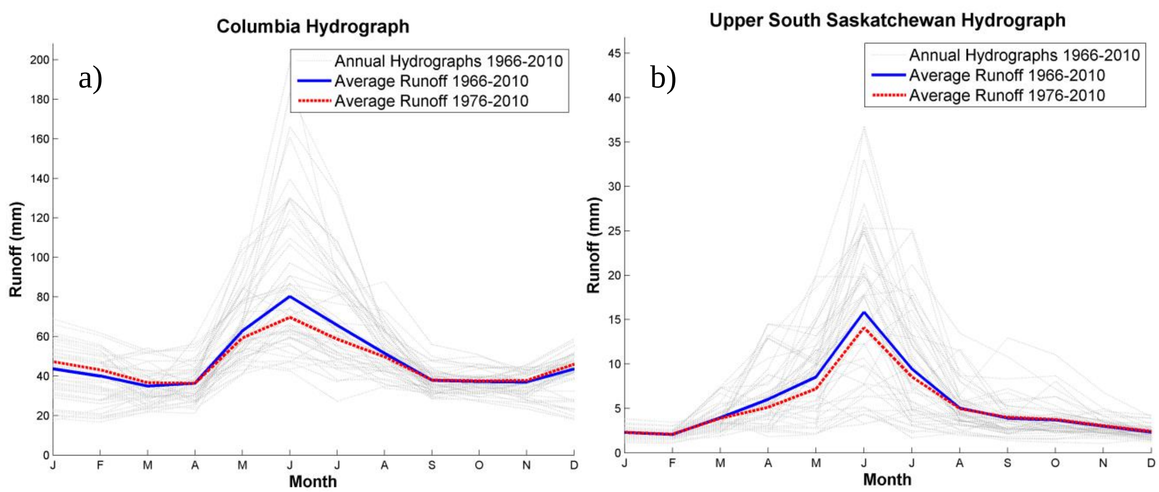


Figure 5.1. Hydrographs for the a) Columbia River basin and b) Upper South Saskatchewan River basin. Average runoff is shown in blue for the period of 1966-2010 and in red for the period of 1976-2010. Annual hydrographs for each individual year (1966-2010) are shown in gray.

5.1.1 Change Point Detection

The Pettitt change point detection test was first applied to the annual runoff series in order to determine whether a) trends in the data could be attributed to change points in the means of the time series rather than long-term, continuous trends, and/or b) a common regime shift (such as the mid-1970s shift) could have been responsible for the majority of trends noted in the CROCWR region. The annual runoff series were analyzed since atmospheric regime shifts tend to occur over the course of a year. Table 5.1 specifies the basins that exhibited change points during at least one analysis period, as well as the year and direction of the identified shift. Downward (upward) pointing arrows indicate that the series mean decreased (increased) after the change point.

Table 5.1 shows that less than half of the CROCWR basins experienced a change point in the annual runoff series during at least one of the analysis periods. In addition, a lack of consistency in the year of shift was found among basins, as well as among individual watersheds for the different analysis periods. Figure 5.2 provides an example of the latter. Even in basins located within the same general region, such as the Mackenzie-Great Slave basins or the Athabasca-Smoky River basins, there was inconsistency among the year of shift. This suggests that the change points that were observed were more likely results of localized changes in one or more controlling factor(s) of runoff in each basin, such as near-surface climate or land-use, rather than a large-scale atmospheric regime shift affecting all or part of the CROCWR study area. It is worth mentioning that, for the 1966-2010 period, one watershed, the Upper South Saskatchewan River basin, experienced a change point around the time of the 1976/1977 regime shift from the negative to positive phase of the PDO; this shift was associated with reduced runoff in the basin.

Table 5.1. Pettitt change point detection results for the CROCWR study area. Watersheds in which a change point was detected during at least one of the analysis periods are listed, along with the year at which the regime shift occurred and the direction (of mean average runoff) of the shift.

Watershed		1976-2010	1966-2010
5	Smoky River	1992 ↓	1984 ↓
7	Lower Athabasca	1999 ↓	1997 ↓
10	West Lake Athabasca	1998 ↓	1996 ↓
12	Great Slave	1989 ↑	1981 ↑*
13	Upper Mackenzie	1997 ↑	
14	Mid Mackenzie		1982 ↓*
15	Lower Mackenzie	1994 ↑	1987 ↑*
24	Upper South Saskatchewan	1990 ↑	1976 ↓
25	Lower South Saskatchewan	1998 ↑	

*More than 4 years of missing data

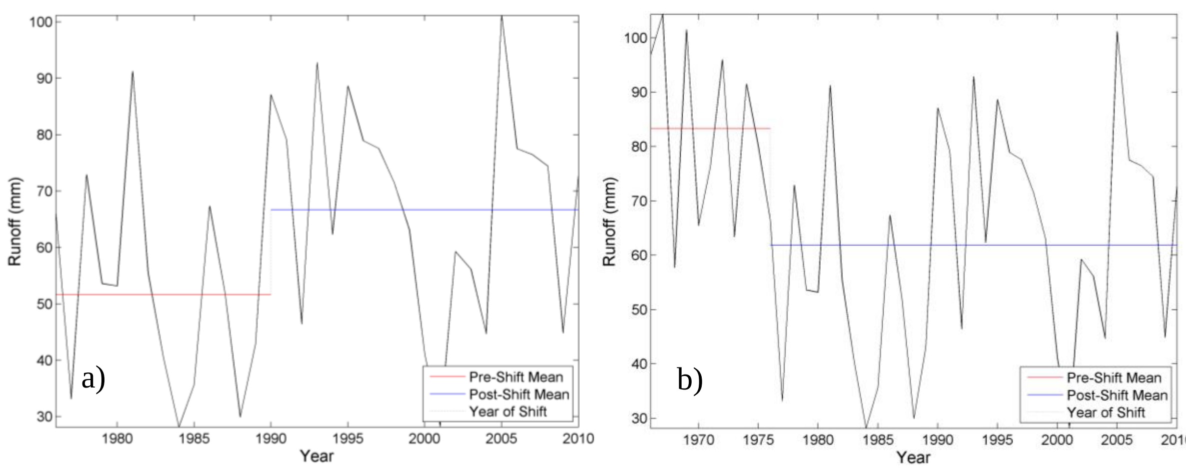


Figure 5.2. Time series plots of runoff in the Upper South Saskatchewan basin for the a) 35- and b) 45-year analysis periods. The pre-shift series means are given by the red lines, while the series means following the detected Pettitt change points are given by the blue lines. The 35-year change point occurred in 1990 and resulted in increased runoff, while the 45-year change point occurred in 1976 and resulted in reduced runoff.

The fact that the year of change point for one basin generally differed with analysis period indicates that the shifts were likely not strong enough to be solely responsible for the trends detected in the data, but instead speaks to the sensitivity of the Pettitt test. Despite differing shift years for each basin, however, change points tended to be in the same direction for all analysis periods in which a shift was detected (with the exception of the South Saskatchewan River basin). In addition, MK trends in basins that did exhibit a shift tended to be significant at the 10% significance level or better in the same direction as the shift.

Due to the lack of consistency among the years of change points, as well as to the minority of basins that exhibited significant change points in the annual data and the fact that these basins were not all concentrated in one region of the CROCWR study area, trend results are considered to represent actual (linear, monotonic) trends in the data and are not considered to reflect single shifts in the series means.

5.1.2 Runoff Trend Results

35-year results

Table 5.2 lists runoff trend slope results in units of mm/yr for the 1976-2010 (35-year) analysis period. Refer to Appendix B, Table B.1 for results in units of mm per 35 year period. Non-significant decreasing (increasing) tendencies are represented by single downward (upward) pointing arrows. Double bolded downward/ upward pointing arrows represent trends significant at the 10% significance level, while double bolded arrows in red indicate trends significant at the 5% significance level. For the 1976-2010 analysis period, winter, fall and warm season runoff trends were determined to be field significant at the 5% significance level, while summer runoff was found to be field significant at the 10% significance level. Trend maps for the 35-year analysis period for each runoff variable are given in Figures 5.3 and 5.5 through 5.7. Time series of annual runoff for two basins that exhibited trends significant at the 5% level are shown in Figure 5.4.

Table 5.2. Runoff trend results for the 1976-2010 (35-year) period.

Watershed	Annual (mm/yr)		Winter++ (mm/yr)		Spring (mm/yr)		Summer+ (mm/yr)		Fall++ (mm/yr)		Cold Season (mm/yr)		Warm Season++ (mm/yr)	
	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend
Upper Liard	0.9	↑	0.2	↑↑	0.6	↑	0	None	0.3	↑	0.4	↑↑	0.3	↑
Fort Nelson	-1.8	↓	0.1	↑↑	-0.6	↓	-1.3	↓	0	None	0.1	↑	-2.4	↓
Lower Liard	2.1	↑↑	0.1	↑	0.7	↑	0.6	↑↑	0.4	↑↑	0.2	↑	1.4	↑
Upper Peace	-0.6	↓	0.6	↑	-0.5	↓	-0.7	↓	-0.5	↓	0.4	↑	-2.0	↓↓
Smoky River	-1.6	↓↓	0	None	-0.8	↓	-1.0	↓↓	0	None	-0.1	↓	-2.1	↓↓
Lower Peace	-0.7	↓	-0.3	↓↓	0.1	↑	-0.1	↓	-0.3	↓	-0.7	↓	-0.5	↓
Upper Athabasca	-0.7	↓	0.1	↑	0.1	↑	-0.6	↓	0.1	↑	0.1	↑	-1.2	↓
Lower Athabasca	-1.7	↓↓	-0.1	↓↓	-0.4	↓	-0.8	↓↓		↓↓	-0.4	↓↓	-1.4	↓↓
East Lake Athabasca	-1.2	↓	-0.1	↓	-0.4	↓	-0.6	↓	-0.2	↓	-0.6	↓	-1.4	↓
West Lake Athabasca	-0.8	↓↓	0	None	-0.2	↓	-0.3	↓	-0.1	↓	-0.1	↓	-0.7	↓↓
Hay	0.2	↑	0	None	-0.1	↓	0	None	0	None	0.1	↑	-0.6	↓
Great Slave	1.6	↑↑	0.4	↑↑	0.7	↑↑	0.3	↑	0.2	↑	0.7	↑↑	0.3	↑
Upper Mackenzie	2.4	↑↑	0	None	1.1	↑↑	1.2	↑↑	0.2	↑	-0.2	↓	2.2	↑↑
Mid Mackenzie	-0.5	↓	0.1	↑	-0.2	↓	-0.4	↓	0.2	↑	0	None	-0.6	↓
Lower Mackenzie	8.3	↑↑	1.0	↑	1.6	↑↑	3.6	↑↑	2.5	↑↑	1.7	↑	4.2	↑
Peel	-0.3	↓	0.2	↑↑	-0.9	↓	0.4	↑	0.4	↑↑	0.3	↑↑	-0.7	↓
North Pacific	1.4	↑	0.2	↑↑	1.2	↑	0.4	↑	-0.3	↓	0.1	↑	0.3	↑
South Pacific	3.2	↑	0.2	↑	1.2	↑	0.3	↑	1.0	↑	0.3	↑	-0.3	↓
Fraser	-0.7	↓	-0.1	↓	0.1	↑	-0.5	↓	0	None	-0.1	↓	-0.8	↓
Okanagan	-0.4	↓	-0.1	↓	0	None	-0.1	↓	-0.1	↓↓	-0.3	↓	-0.5	↓
Columbia	-1.2	↓	-0.9	↓↓	0.1	↑	0	None	-0.2	↓	-1.5	↓↓	-0.3	↓
Upper North Saskatchewan	-0.7	↓	-0.1	↓↓	-0.1	↓	-0.3	↓	-0.1	↓	-0.2	↓	-0.6	↓
Lower North Saskatchewan	0	None	0	None	0	None	0	None	0	None	0.1	↑	0	None
Upper South Saskatchewan	0.4	↑	0	None	0.2	↑	0.1	↑	0.1	↑↑	0	None	0.3	↑
Lower South Saskatchewan	0.5	↑	0	None	0.1	↑	0.3	↑↑	0.2	↑↑	0.3	↑	0.2	↑

Table 5.2(con't). Runoff trend results for the 1976-2010 (35-year) period.

Station	Annual		Winter++		Spring		Summer+		Fall++		Cold Season		Warm Season++	
	#	%	#	%	#	%	#	%	#	%	#	%	#	%
# ↓	11	44%	3	12%	10	40%	10	40%	7	28%	8	32%	12	48%
# **↓↓**	3	12%	4	16%	0	0%	2	8%	2	8%	2	8%	4	16%
# ↑	6	24%	6	24%	10	40%	5	20%	6	24%	10	40%	7	28%
# **↑↑**	4	16%	5	20%	3	12%	4	16%	5	20%	3	12%	1	4%
# No trend	1	4%	7	28%	2	8%	4	16%	5	20%	2	8%	1	4%
# Missing	0	0%	0	0%	0	0%	0	0%	0	0%	0	0%	0	0%

+ Variable is field significant at 10% significance level.

++ Variable is field significant at 5% significance level

↑ or ↓: Non-significant trend

↑↑ or **↓↓**: Trend significant at 10% significance level

↑↑ or ↓↓: Trend significant at 5% significance level

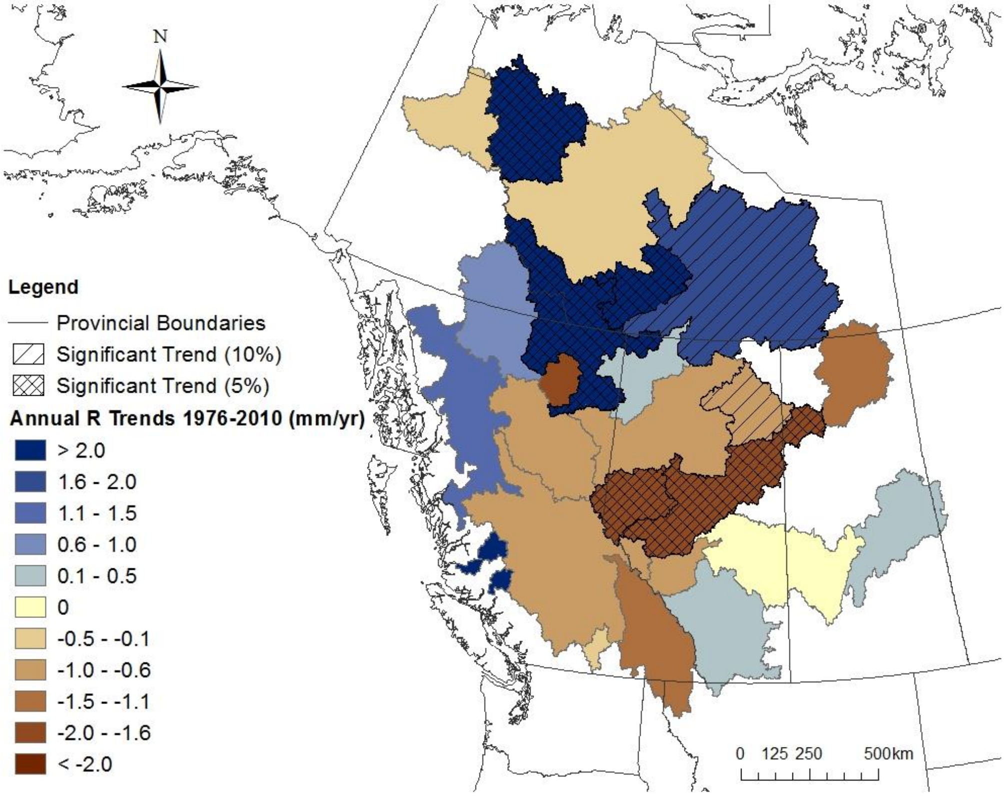


Figure 5.3. Map of annual runoff trend slopes for the 1976-2010 analysis period. Basins exhibiting significant trends are shown as hatched.

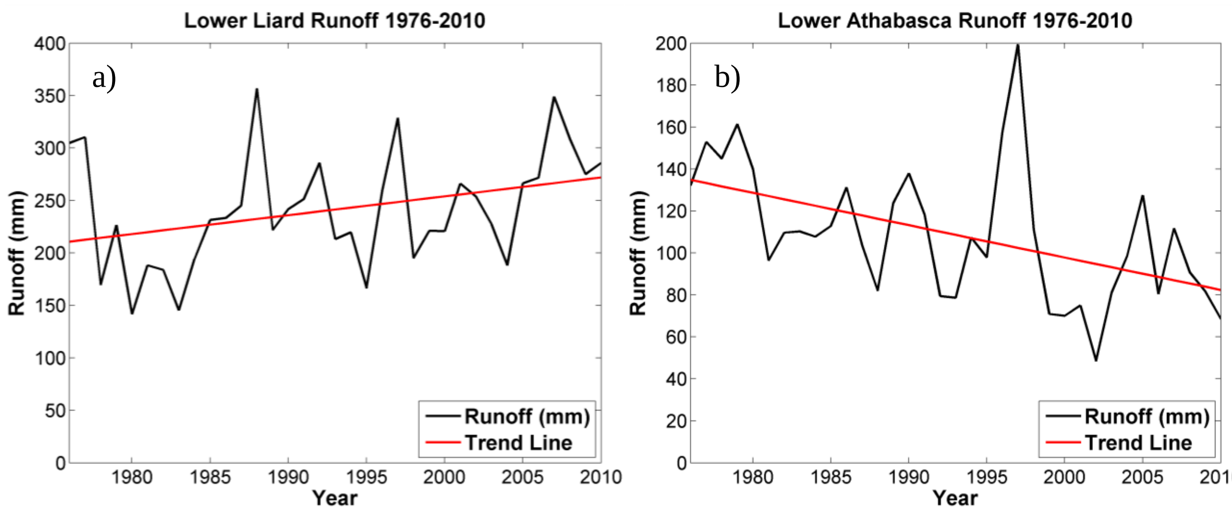
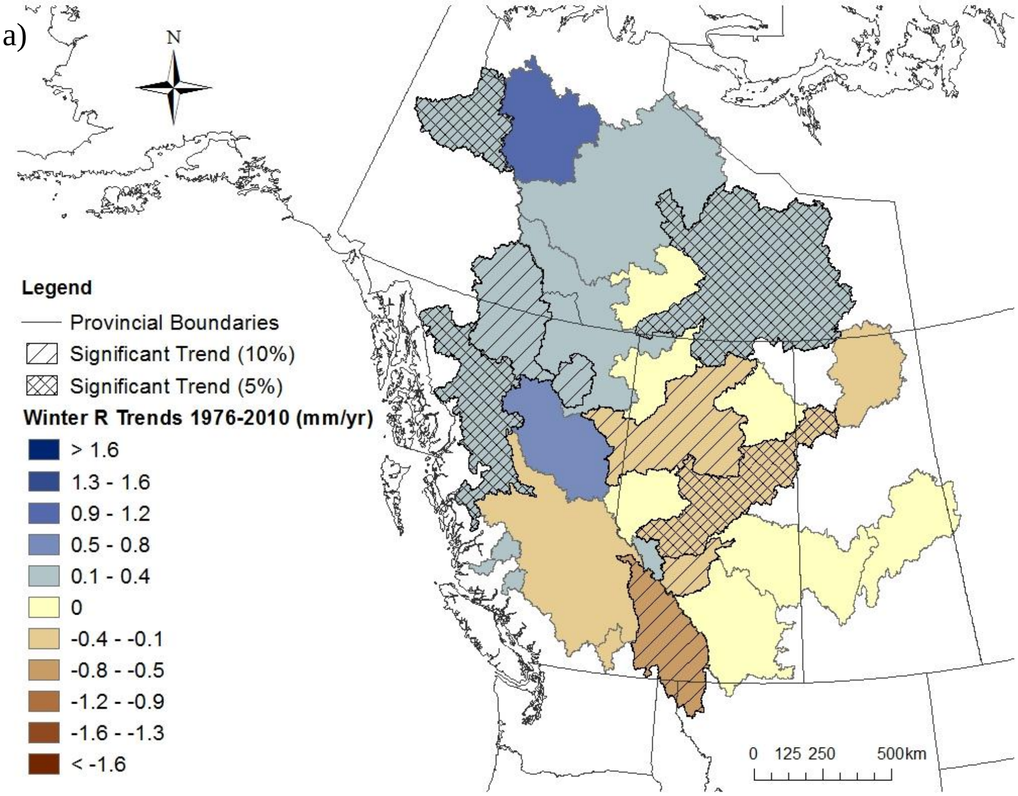


Figure 5.4. Time series of 1976-2010 annual runoff for the a) Lower Liard and b) Lower Athabasca watersheds. The Lower Liard exhibited an increasing trend of 2.1 mm/yr (75.7 mm/35yrs), while the Lower Athabasca displayed a decreasing trend of -1.7 mm/yr (-60.3 mm/35yrs), both significant at the 5% significance level.

a)



b)

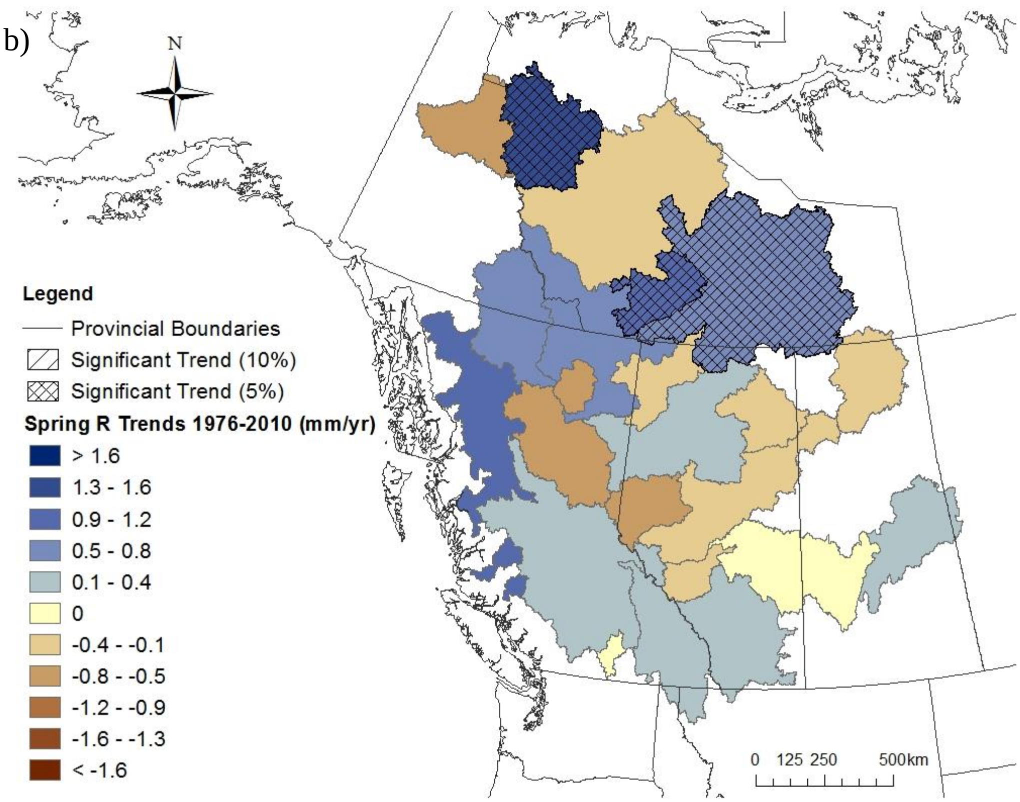


Figure 5.5. Same as Figure 5.3 but for a) winter and b) spring. Note that scale differs from Figure 5.3.

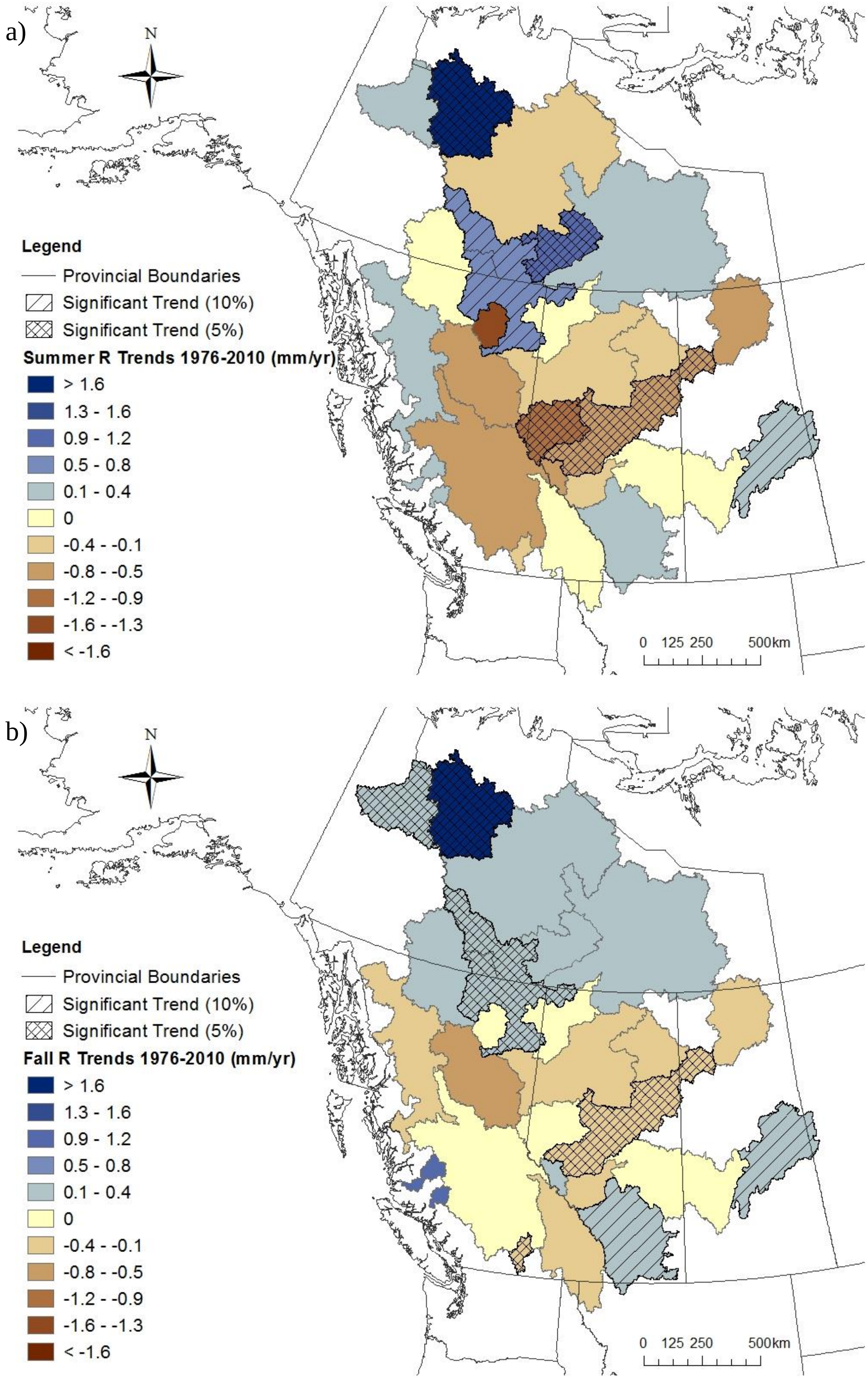


Figure 5.6. Same as Figure 5.3 but for a) summer and b) fall. Note that scale differs from Figure 5.3.

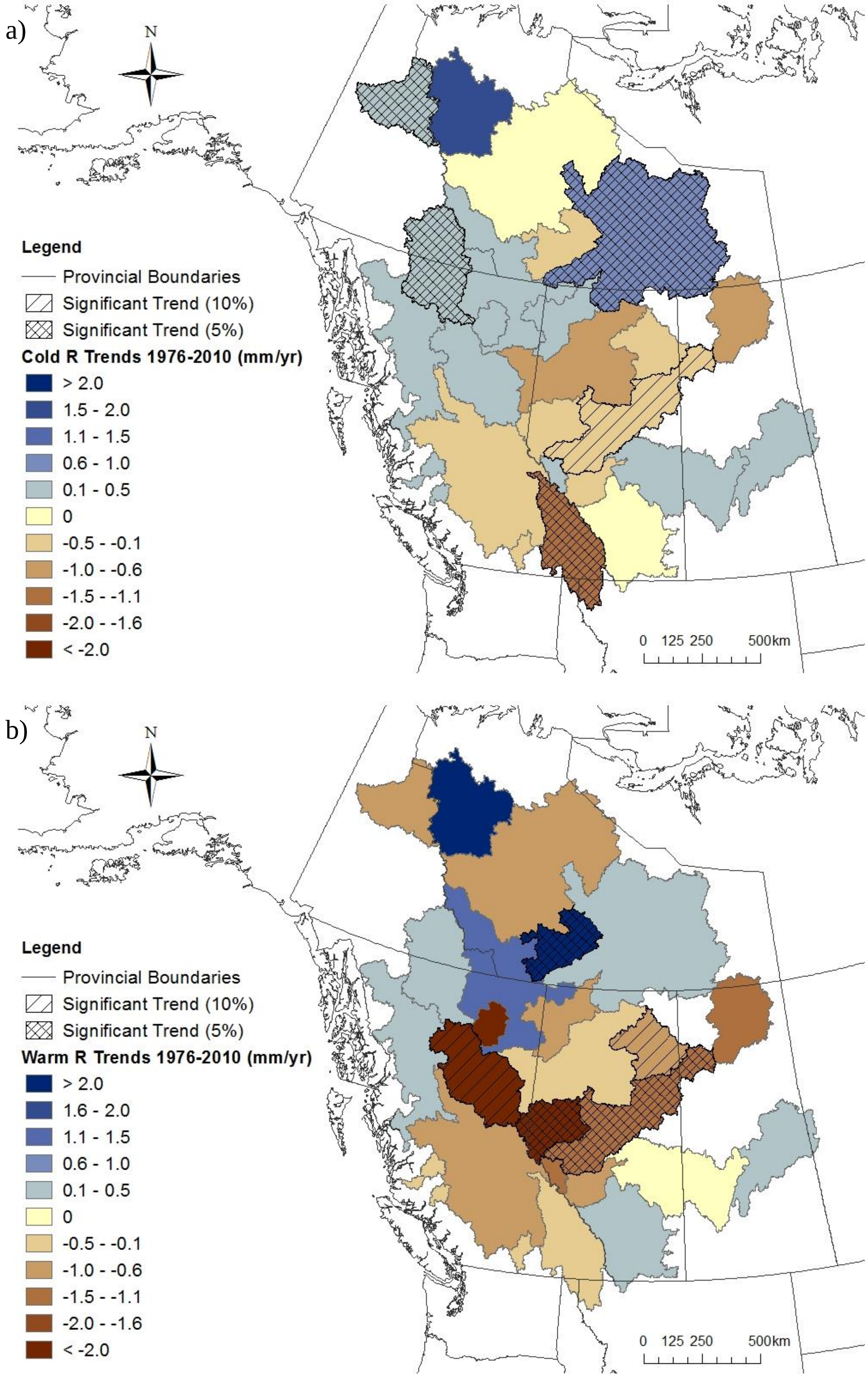


Figure 5.7. Same as Figure 5.3 but for a) cold season and b) warm season runoff. Note that scale differs from Figures 5.5 and 5.6.

For most seasons, MK trend analysis revealed a general tendency towards higher runoff in the most northern basins with contrasting decreases in runoff in watersheds stretching across the mid-latitudes of the study region. This pattern was clearest during the fall and winter seasons, during which all watersheds north of 60°N exhibited increasing (or a lack of) trends in runoff. Winter also saw the largest number of basins with trends significant at the 10% level or better. Trend magnitudes were generally strongest during the summer and warm season due to the fact that overall runoff is significantly lower during the winter months, therefore changes during cold periods/seasons are likely to be smaller.

There are some clear exceptions to the above discussed pattern. The Peel and Mid-Mackenzie basins exhibited slight decreasing tendencies on an annual basis, as well as during the spring, summer (Mid-Mackenzie only), and warm seasons, distinguishing them from the pattern observed for the other watersheds north of 60°N, but in agreement with the results of Déry and Wood (2005). The impact of storage effects may be partly responsible for this observation in the Mid-Mackenzie basin, within which Great Bear Lake is located.

Unlike most other basins located among the mid-latitudes of the study region, the North Pacific basin exhibited increasing trends for all variables (except fall). This is likely explained by the fact that the Pacific coast does not experience the same extremes in temperature as does the interior, therefore a warming (or cooling) climate may not have the same effect on runoff in this area as it would on rivers located in interior western Canada. In addition, since streamflow patterns on the Pacific coast are mostly rainfall-dominated (rather than snowmelt-dominated), changes in snow accumulation and snowfall would not have the same effect here as they would in the interior of BC, Alberta, or the territories.

No consistent trend direction was noted in the southern-most basins in the study region. It was, however, interesting and unexpected to have slight increasing trends in the South Saskatchewan River basins for the majority of variables, especially given the results of previous research which suggested a strong drying of this region (e.g., Schindler and Donahue, 2006). This result may be attributable to the large degree of regulation this basin has undergone (e.g., construction of the Gardiner Dam), and/or to the relatively short period of record studied. Additionally, increasing trends in the Lower South Saskatchewan River basin could be attributable to the 1998 PDO regime shift, when the "famous" dry period that began in the mid-1970s ended and a wet period began. It is worth noting that at least the Upper South Saskatchewan River basin exhibited decreasing (or a lack of) trends for all variables (except fall) for the 45-year analysis period; the Lower South Saskatchewan River basin, on the other hand, showed a mix of increasing or lack of trends during this period for the variables with available data.

Spring runoff trend results differed most from the remaining variables in that the general pattern of increasing runoff in the North and decreasing runoff among the mid-latitudes did not hold. During spring, the largest number of basins exhibited overall increasing trends in runoff, including the majority of the southern-most basins, as well as the Lower Peace River basin located along mid-latitude band. Interestingly, spring was also associated with the greatest number of basins exhibiting decreasing trends in mean temperature (see *Section 5.3.1*).

Given the general trend pattern noted with degree of latitude for most variables, it was desirable to both quantify this relationship, as well as to explore the relationship between runoff and other characteristics of a basin's regime/topography. Correlation analysis was performed between runoff (mean and trend) and latitude, longitude (both taken at the centroid of each watershed), drainage area, and median basin elevation, a crude measure of basin hypsometry. Table 5.3 presents the results of correlation analysis, and confirms the significant relationship between runoff trend and basin latitude. In addition, a relationship between runoff trend and degree of longitude was noted for the fall, winter, and cold seasons, as well as for annual runoff, while mean runoff during the spring, summer, warm season, and annually were shown to be linked to degree longitude. A clear positive relationship was also present between average runoff and watershed elevation, while runoff trend was significantly linked to hypsometry during the summer and warm season. As expected, watershed area had no influence on either runoff or trend, since the influence of area was removed via the calculation of runoff depth from streamflow. Percent glacier coverage over each watershed would be another useful factor to explore, given available data (e.g., Déry *et al.*, 2012).

Table 5.3. Pearson correlation between basin topography and 1976-2010 runoff (mean and trend).

1976-2010 Runoff Characteristic		Latitude	Longitude	Drainage Area	Median Elevation
Annual	Mean Runoff	-0.22	**<u>-0.43</u>**	-0.17	**<u>0.59</u>**
	Runoff Trend	**<u>0.42</u>**	**-0.37**	0.01	-0.27
Winter	Mean Runoff	-0.32	-0.21	-0.07	**<u>0.56</u>**
	Runoff Trend	**<u>0.55</u>**	**-0.42**	0.02	-0.26
Spring	Mean Runoff	-0.18	**<u>-0.53</u>**	-0.13	**<u>0.61</u>**
	Runoff Trend	0.14	-0.24	0.09	-0.07
Summer	Mean Runoff	-0.17	**<u>-0.42</u>**	-0.21	**<u>0.53</u>**
	Runoff Trend	**<u>0.44</u>**	-0.31	0.02	**-0.35**
Fall	Mean Runoff	-0.26	-0.31	-0.14	**<u>0.55</u>**
	Runoff Trend	**<u>0.43</u>**	**-0.39**	-0.04	-0.22
Cold Season	Mean Runoff	**-0.36**	-0.22	-0.10	**<u>0.59</u>**
	Runoff Trend	**<u>0.51</u>**	**-0.39**	0.05	-0.27
Warm Season	Mean Runoff	-0.16	**<u>-0.48</u>**	-0.18	**<u>0.55</u>**
	Runoff Trend	**0.39**	-0.20	0.10	**-0.37**

Note: Significant correlations are **bolded** (10%) and **<u>underlined</u>** (5%).

45-year trend results

Table 5.4 shows runoff trend slope results in units of mm/yr for the 1966-2010 (45-year) analysis period. Refer to Appendix B, Table B.2 for results in units of mm per 45 years. Only winter runoff was found to be field significant at the 5% significance level, while summer and cold season runoff were determined to be field significant at the 10% significance level. 45-year trend maps are located in Appendix B, Figures B.1 through B.4.

For the 45-year analysis period, the majority of the northern-most basins lacked sufficient data for trend analysis; the mid- and lower-latitude watersheds, however, showed trend patterns and magnitudes similar to those found for the 35-year period. Some differences worth noting include the general tendency for much stronger magnitude decreasing trends during the warm season for the 45-year period, as well as a lack of increasing trends in the southern-most basins during spring. In addition and as previously mentioned, the Upper South Saskatchewan River basin only exhibited increased runoff during the fall. Overall, drying in the lower latitudes is shown to have been stronger over the 45-year period, particularly in the southern-most basins.

This result has important implications concerning the mid-1970s shift in the PDO. Although no basin (with the exception of the Upper South Saskatchewan River) exhibited a significant change point around the time of the 1976/1977 PDO phase shift, it is likely that runoff was still influenced overall by this change. In 1976/1977, the PDO changed from a cool, wet phase associated with higher than normal runoff to a warm, dryer phase associated with lower than normal runoff. Given that decreasing trends in the mid- and especially lower-latitude basins (almost all of which are sourced by the Rocky Mountains) were stronger over the longer time period suggests that trend magnitudes were influenced by the presence of the shift (whether or not a one-time shift was detected). However, the fact that this drying trend was also noted, especially among the mid-latitudes, strictly post-1976 shift indicates a drying tendency not related to the PDO. Furthermore, the PDO has been reported to have shifted back to a wet phase in 1998, as mentioned, but only the Lower South Saskatchewan River basin displayed a significant shift in the direction of increased runoff related to this shift (for the shorter analysis period).

It appears, therefore, that the overall drying trend in the mid-latitude watersheds sourced by the Rocky Mountains were related in part to the PDO, but that some change (i.e., in the climatic system) outside of the PDO and spanning both of its phases was also responsible for these trends, especially for the shorter analysis period. Consistency in the direction of trend was not as clear in the most southern watersheds, at least for the shorter analysis period.

Table 5.4. Runoff trend results for the 1966-2010 (45-year) period.

Watershed	Annual (mm/yr)		Winter[++] (mm/yr)		Spring (mm/yr)		Summer[+] (mm/yr)		Fall (mm/yr)		Cold Season[+] (mm/yr)		Warm Season (mm/yr)	
	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend
Upper Liard	0.4	↑	0.2	↑↑	0.3	↑	-0.2	↓	0.2	↑↑	0.5	↑↑	0.2	↑
Fort Nelson	-0.3	↓	0.1	↑↑	-0.2	↓	-0.5	↓	0.1	↑	0.3	↑↑	-0.3	↓
Lower Liard	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Upper Peace	-0.1	↓	1.2	↑↑	-0.6	↓	-0.3	↓	0.6	↑	2.1	↑	-1.3	↓
Smoky River	-1.4	↓↓	0	None	-1.0	↓↓	-0.4	↓↓	0	None	-0.1	↓	-1.4	↓↓
Lower Peace	-0.1	↓	-0.1	↓	0.3	↑	-0.2	↓	-0.2	↓↓	-0.3	↓	-0.1	↓
Upper Athabasca	-1.3	↓	0.1	↑	-0.3	↓	-0.9	↓	0.1	↑	0.1	↑	-1.3	↓
Lower Athabasca	-1.5	↓↓	-0.1	↓↓	-0.4	↓↓	-0.5	↓↓	-0.2	↓↓	-0.2	↓	-1.1	↓↓
East Lake Athabasca	0.2	↑	0.1	↑	0	None	-0.1	↓	0	None	0.1	↑	-0.3	↓
West Lake Athabasca	-0.6	↓↓	0	None	-0.2	↓↓	-0.2	↓	-0.1	↓	0	None	-0.4	↓
Hay	0.6	↑	0	None	0	None	0.3	↑	0.1	↑↑	0.1	↑↑	0.3	↑
Great Slave	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Upper Mackenzie	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mid Mackenzie	-	-	-	-	-	-	-0.5	↓↓	-	-	-	-	-	-
Lower Mackenzie	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Peel	-	-	-	-	-	-	-	-	-	-	-	-	-	-
North Pacific	0.8	↑	0.4	↑↑	0.9	↑	-0.2	↓	0.0	None	0.8	↑	1.2	↑
South Pacific	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Fraser	-1.2	↓↓	-0.1	↓	-0.2	↓	-0.6	↓↓	-0.2	↓	0	None	-1.1	↓↓
Okanagan	-0.1	↓	0	None	0	None	0.1	↑	0	None	-0.1	↓	0.1	↑
Columbia	-1.8	↓	0.5	↑	-1.2	↓↓	-1.0	↓	0.3	↑	0.7	↑	-2.3	↓
Upper North Saskatchewan	-0.7	↓	0.1	↑	-0.4	↓	-0.6	↓	0.2	↑↑	0.4	↑↑	-1.1	↓↓
Lower North Saskatchewan	0	None	0	None	0	None	0	None	0	None	0	None	0	None
Upper South Saskatchewan	-0.4	↓	0	None	-0.3	↓	-0.1	↓	0.1	↑↑	-0.1	↓	-0.2	↓
Lower South Saskatchewan	-	-	-	-	0	None	0.1	↑	0.1	↑	-	-	0	None

Table 5.4(con't). Runoff trend results for the 1966-2010 (45-year) period.

Station	Annual		Winter++		Spring		Summer+		Fall		Cold Season+		Warm Season	
	#	%	#	%	#	%	#	%	#	%	#	%	#	%
# ↓	8	**32%**	2	**8%**	6	**24%**	11	**44%**	2	**8%**	5	**20%**	8	**32%**
# ↓↓	4	**16%**	1	**4%**	4	**16%**	4	**16%**	2	**8%**	0	**0%**	4	**16%**
# ↑	4	**16%**	4	**16%**	3	**12%**	3	**12%**	5	**20%**	5	**20%**	4	**16%**
# ↑↑	0	**0%**	4	**16%**	0	**0%**	0	**0%**	4	**16%**	4	**16%**	0	**0%**
# No trend	1	**4%**	6	**24%**	5	**20%**	1	**4%**	5	**20%**	3	**12%**	2	**8%**
# Missing	8	**32%**	8	**32%**	7	**28%**	6	**24%**	7	**28%**	8	**32%**	7	**28%**

+ Variable is field significant at 10% significance level.

++ Variable is field significant at 5% significance level.

↑ or ↓: Non-significant trend

↑↑ or **↓↓**: Trend significant at 10% significance level

↑↑ or ↓↓: Trend significant at 5% significance level

5.1.3 Freshet Trend Results

For the streamflow timing variables, the 10P date was shown to less effectively capture the onset of the spring freshet for a number of the CROCWR stations, therefore only the pulse date results are presented as measurement of the timing of the start of the spring freshet. Figure 5.8 shows two examples of stations at which the 10P date measurement failed to capture this significant hydrological event.

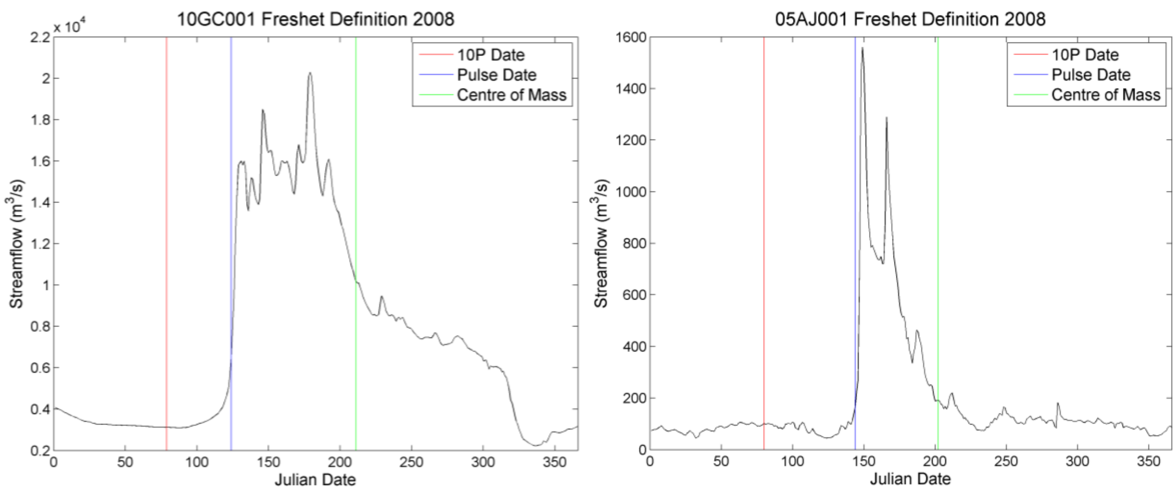


Figure 5.8. Daily flows for a) Mackenzie River at Fort Simpson (WSC station 10GC001) and b) South Saskatchewan River at Medicine Hat (WSC station 05AJ001) for the year 2008. The 10P dates are shown by the red lines (Julian days 79 and 80, respectively), whereas the spring pulse dates are shown by the blue lines (Julian days 124 and 144, respectively). The CM dates (using the pulse dates as the starting dates) are shown by the green lines.

Due to regulation of the Peace River, the spring freshet no longer occurs naturally at the Peace River station near Taylor (WSC station 07FD002), located approximately 125 km downstream of the W.A.C. Bennett Dam. Figure 5.9 shows the effect that the dam has had on upper Peace River streamflow since its 1968 installation.

Figure 5.9 shows that installation of the Bennett Dam has essentially flattened the hydrograph of the upper Peace River. As shown in Figure 5.10, the 10P, pulse, and CM dates could not be accurately identified, therefore this station was left out of the analysis. Note that although a number of other CROCWR stations are also regulated, the Peace River station near Taylor is the only one at which the hydrograph has been completely altered, hence all other stations were retained in the analysis.

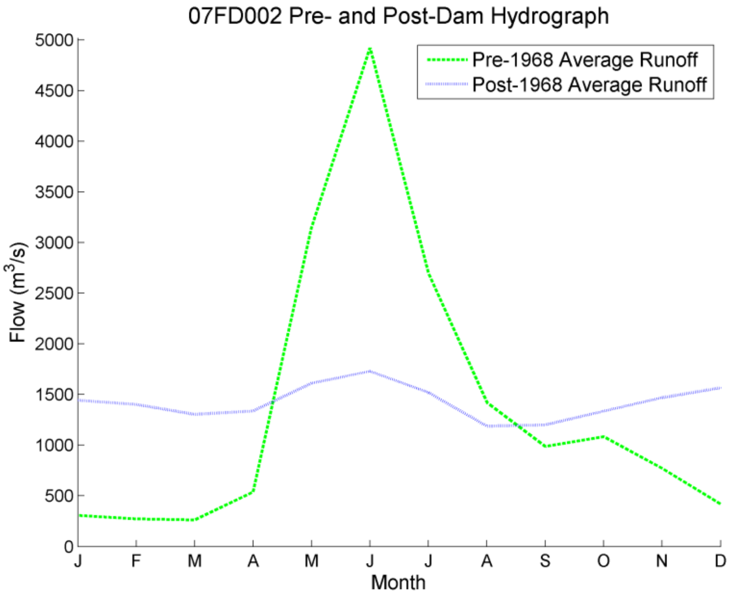


Figure 5.9. Average streamflow hydrographs for the Peace River station near Taylor (WSC station 07FD002) before (1961-1967) and after (1968-2010) the 1968 installation of the W.A.C. Bennett Dam. Pre-dam streamflow is shown in green, while post-dam streamflow is shown in blue.

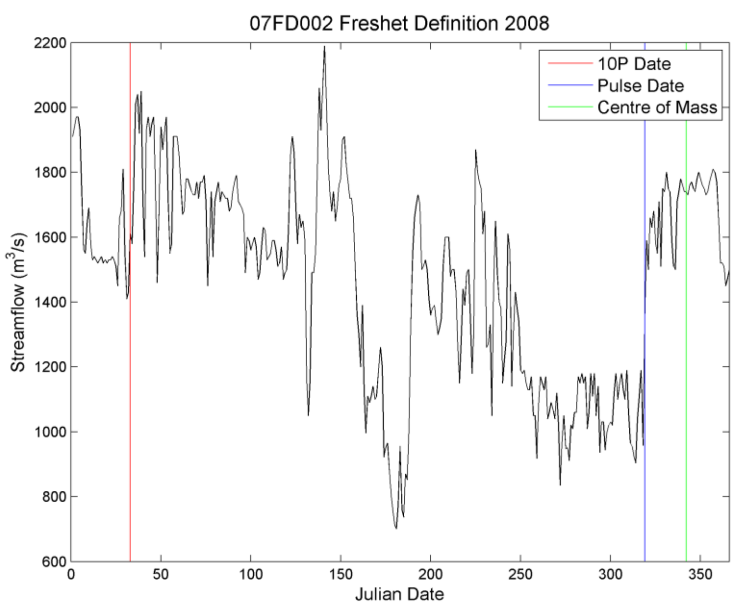


Figure 5.10. Daily flows for the Peace River near Taylor (WSC station 07FD002) for the year 2008. The 10P date is shown by the red line (Julian day 33 or February 2), the spring pulse date is shown by the blue line (Julian day 319 or November 13), and the CM date (using the pulse date as the starting date) is shown by the green line.

35-year results

Table 5.5 shows freshet trend results for the 1976-2010 (35-year) period. Pulse date, CM, and length of freshet slopes are given in units of days/35yrs, while trend slopes for freshet volume (measured as runoff) are given in units of mm/35yrs. Trend direction and significance are given in the same format as runoff trends presented in Table 5.2. No freshet measurement was found to be field significant at either the 5% or 10% significance level for the 35-year period. 1976-2010 trend maps for each freshet variable are given in Figures 5.11 and 5.12. Time series of freshet data for one station that exhibited significant trends in two freshet variables are shown in Figure 5.13.

Stations showed a mix of both increasing and decreasing trends in pulse date for the 35-year period. At the northern-most stations as well as along the Pacific coast, a general tendency toward earlier onset of the spring freshet (i.e., decreasing pulse date trends) was noted. At least in the more northern rivers, this is likely related to a shift toward earlier snowmelt from May to April, as identified by Linton (2013). At the more southern latitudes, increasing trends indicating later pulse dates were noted among the Rocky Mountain headwater stations, likely causing the increasing trends in spring runoff noted in *Section 5.1.2*. Conversely, decreasing pulse date trends were noted for the Lower South Saskatchewan River basin. CM dates mainly decreased (non-significantly) throughout the entire study region, indicating an earlier end to the spring freshet.

Freshet length trends were more influenced by the magnitude of the pulse date trends versus the CM date trends, as pulse dates were generally much stronger (i.e., a greater number of days earlier per 35 years). In the northern-most basins, the length (in days) of the spring freshet was shown to have significantly increased over the 35-year period, whereas shorter spring freshets were observed in the majority of the more southern latitude rivers, including along the northern Pacific coast. Consequently, rivers in the North (with the exception of the Peel River) and along the Pacific Coast experienced increased freshet volumes, whereas gauging stations in the mid-latitudes saw mostly decreased freshet runoff. In the southern latitudes, a mix of both increased and decreased freshet volumes were detected. These results are in line with the results of 35-year warm season runoff trend analysis. Decreases in mid- and low-latitude river freshet volumes are likely related in part to decreases in overall snow accumulation (Linton, 2013), whereas increases in northern latitude river freshet volumes are likely related more to changes in other climatic factors, including temperature and/or precipitation (see *Section 5.3.1*).

Table 5.5. Freshet trend results for the 1976-2010 (35-year) period.

Station	Pulse Date (days/35yrs)		CM Date (days/35yrs)		Length (days/35yrs)		Volume (mm/35yrs)	
	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend
10BE001	-2	↓	0	None	0	None	0.1	↑
10CD001	-14	↓↓	-1	↓	16	↑↑	-0.3	↓
10ED002	-8	↓↓	0	None	8	↑↑	0.3	↑
07FD002	-	-	-	-	-	-	-	-
07GJ001	2	↑	-2	↓	-8	↓	-0.4	↓
07KC001	-10	↓↓	-6	↓↓	-8	↓	-0.4	↓↓
07AD002	7	↑	0	None	-5	↓	0	None
07DA001	0	None	-6	↓	-9	↓	-0.3	↓↓
07LE002	0	None	13	↑	30	↑	-0.1	↓
07JD002	0	None	0	None	-12	↓	-0.2	↓
07OB001	12	↑	-1	↓	-8	↓	-0.1	↓
07NB001	-3	↓	-2	↓	0	None	0	None
FortProvidence	46	↑↑	26	↑↑	-36	↓↓	0.1	↑
10GC001	13	↑	6	↑	-22	↓↓	0.1	↑
10KA001	-39	↓	-15	↓	36	↑	0.1	↑
10LC014	0	None	17	↑	-4	↓	0	None
10MC002	-6	↓	-2	↓	3	↑	0	None
08BB005	25	↑	0	None	-12	↓	-0.1	↓
08CE001	0	None	0	None	-3	↓	0	None
08CG001	-3	↓	0	None	0	None	-0.1	↓
08DB001	-4	↓	-3	↓	2	↑	-0.1	↓
08EF001	-5	↓↓	-4	↓	2	↑	0	None
08FF001	-5	↓↓	-4	↓	5	↑↑	0.1	↑
08FA002	-10	↓	-3	↓	2	↑	0	None
08FB006	-7	↓	0	None	7	↑	0.1	↑
08FB007	-8	↓	9	↑	19	↑↑	-0.4	↓
08FC003	5	↑	2	↑	-2	↓	1.1	↑
08GD004	5	↑	-2	↓	-8	↓↓	0	None
08MF005	-2	↓	-7	↓	-4	↓	0.9	↑
08NM085	0	None	-2	↓	-6	↓	0.5	↑
08NE058	-4	↓	-4	↓	-1	↓	0.4	↑↑
05DF001	0	None	0	None	-7	↓	0.6	↑
05GG001	-5	↓	0	None	11	↑	1.1	↑↑
05AJ001	-2	↓	-1	↓	-6	↓	-0.2	↓
05CK004	3	↑	-2	↓	-12	↓	0	None
05HG001	-5	↓	6	↑	9	↑	-0.2	↓
05KJ001	-9	↓	0	None	10	↑↑	0.7	↑↑

Table 5.5 (con't). Freshet trend results for the 1976-2010 (35-year) period.

Station	Pulse Date		CM Date		Length		Volume	
	#	%	#	%	#	%	#	%
# ↓	20	56%	18	50%	19	53%	13	36%
# **↓↓**	5	14%	1	3%	3	8%	2	6%
# ↑	9	25%	7	19%	14	39%	14	39%
# **↑↑**	1	3%	1	3%	5	14%	3	8%
# No trend	7	19%	11	31%	3	8%	9	25%
# Missing	1	3%	1	3%	1	3%	1	3%

↑ or ↓: Non-significant trend

↑↑ or **↓↓**: Trend significant at 10% significance level

↑↑ or **↓↓**: Trend significant at 5% significance level

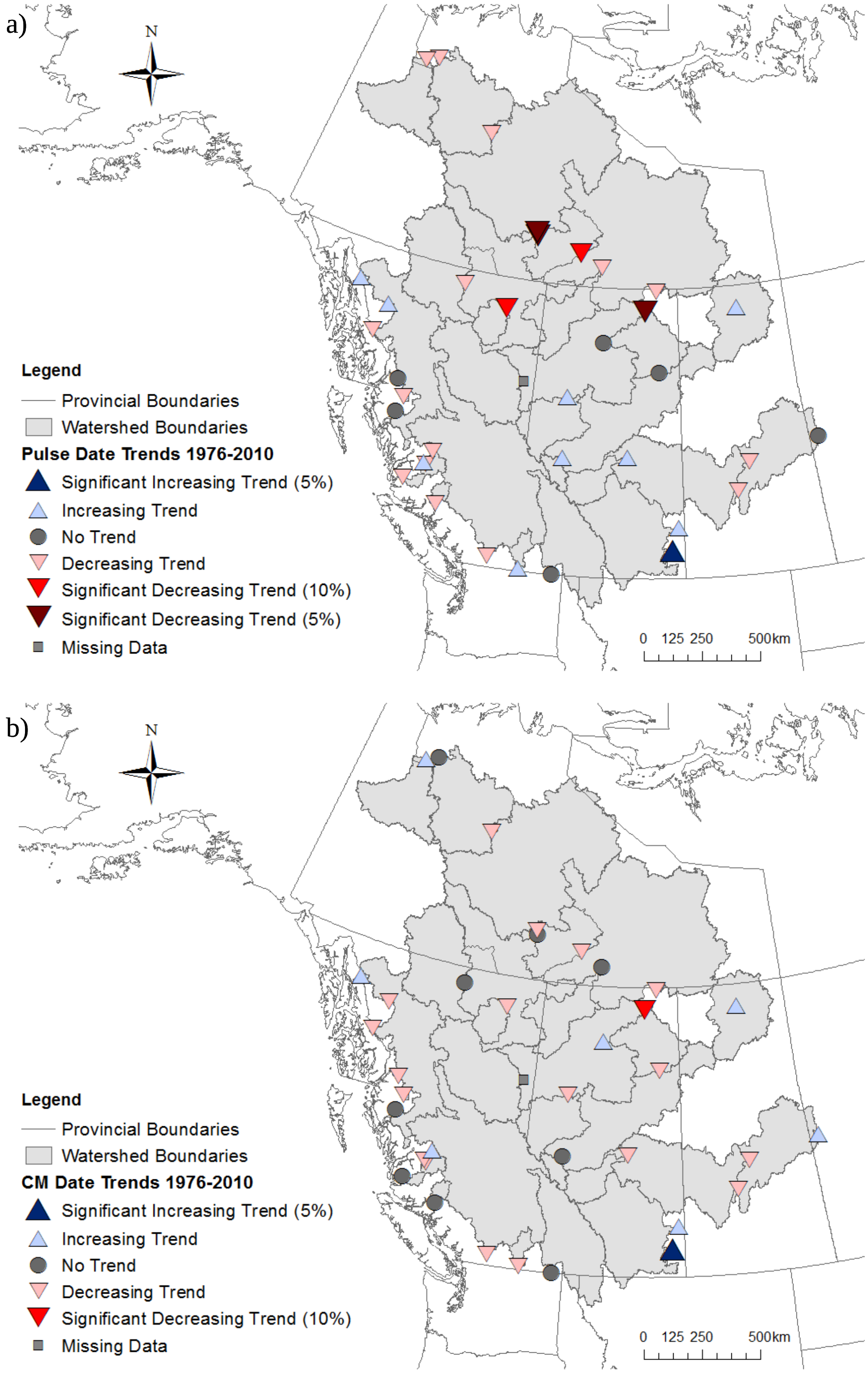


Figure 5.11. Map of a) pulse date and b) CM date trends for the 1976-2010 analysis period.

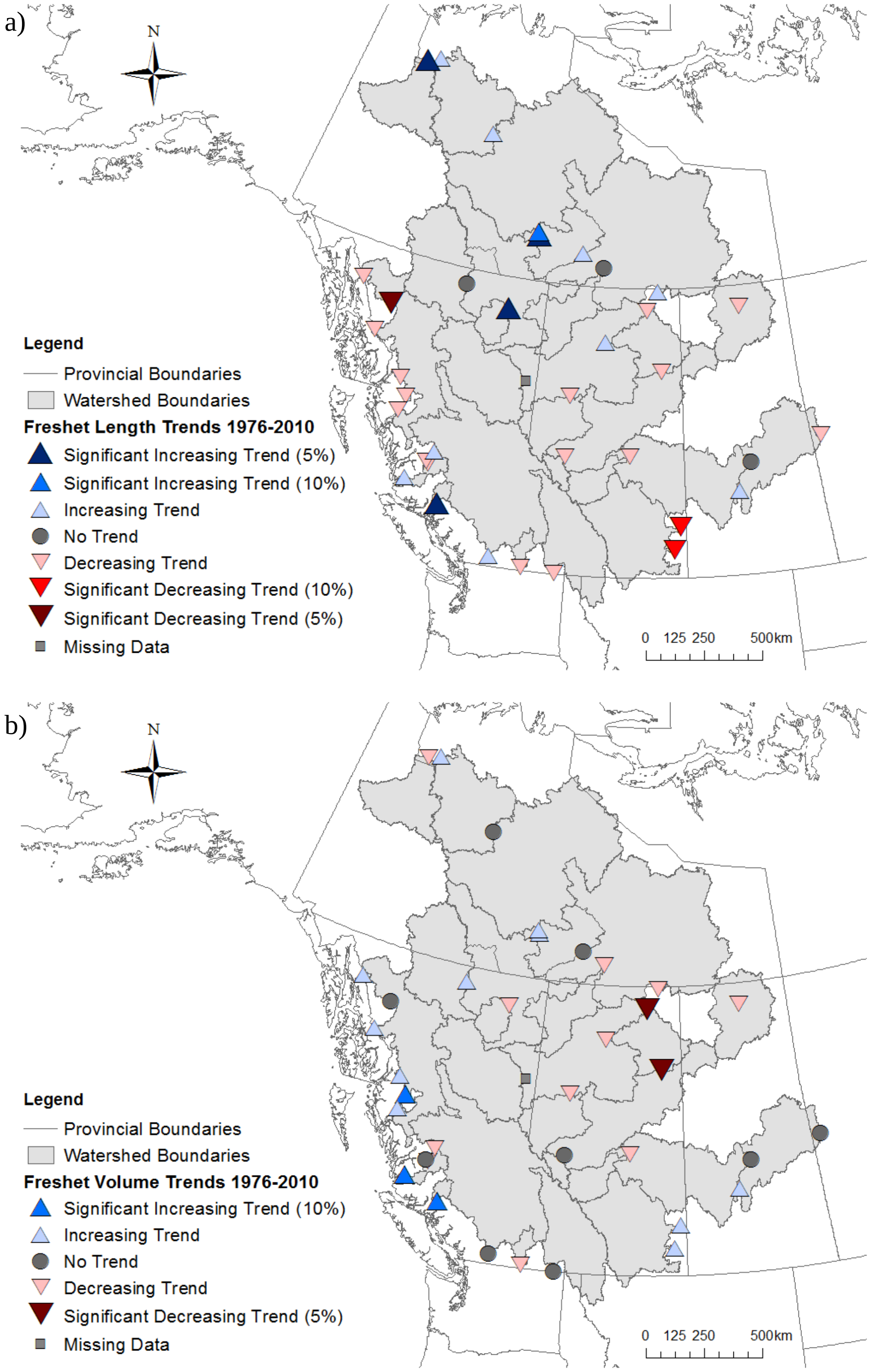


Figure 5.12. Map of a) freshet length and b) freshet volume (measured as runoff) trends for the 1976-2010 analysis period.

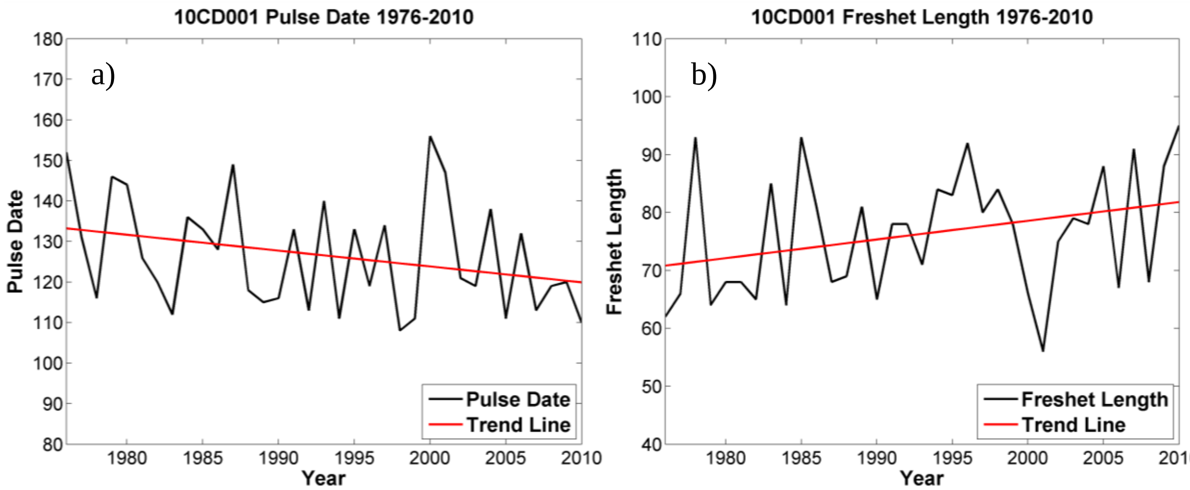


Figure 5.13. Time series of a) pulse date and b) freshet length data for the Muskwa River near Fort Nelson station (WSC station 10CD001). The pulse date exhibited a trend of -14 days/35yrs (10% significant), while the freshet length exhibited a trend of +16 days/35yrs (5% significant).

45-year results

Table 5.6 shows freshet trend results for the 1966-2010 (45-year) period. Pulse date, CM, and length of freshet slopes are given in units of days/45yrs, while trend slopes for freshet volume (measured as runoff) are given in units of mm/45yrs. Pulse date and CM date were found to be field significant at 5%, while freshet length and volume were found to be field significant at the 10% level. Trend maps for the 45-year period are located in Appendix B, Figures B.5 and B.6.

45-year pulse date trends were consistent with the 35-year results, in that mainly decreasing tendencies (i.e., earlier spring freshets) were observed among the northern-most and Pacific coast rivers, while more increasing trends (i.e., later pulse dates) were noted in the interior and south of the study region (with the exception of the Lower South Saskatchewan River basin, where significant decreasing trends were observed).

For the CM date and freshet length, contrasting observations were made between 45- and 35-year observations. For the 45-year analysis period, mostly increasing trends and tendencies in CM dates were noted for the majority of stations with sufficient data, indicating later termination of the spring freshet. Primarily longer spring freshets were noted overall as well. It is therefore interesting that mostly decreased freshet volumes (runoff) were noted at the mid- and lowest-latitude stations, although this was in agreement with 45-year warm season runoff trend results. Mainly increasing freshet volumes were noted at the highest latitude stations (with available data).

Table 5.6. Freshet trends results for the 1966-2010 (45-year) period.

Station	Pulse Date[++] (days/45yrs)		Centre of Mass[++] (days/45yrs)		Length[+] (days/45yrs)		Volume[+] (mm/45yrs)	
	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend
10BE001	-2	↓	1	↑	2	↑	0.1	↑
10CD001	-6	↓	3	↑	8	↑	-0.1	↓
10ED002	-	-	-	-	-	-	-	-
07FD002	-	-	-	-	-	-	-	-
07GJ001	5	↑	7	↑	0	None	-0.5	↓↓
07KC001	-13	↓↓	9	↑	9	↑	-0.5	↓↓
07AD002	3	↑	1	↑	-1	↓	-0.2	↓
07DA001	0	None	0	None	0	None	-0.3	↓↓
07LE002	-	-	-	-	-	-	-	-
07JD002	0	None	19	↑	27	↑	-0.1	↓
07OB001	-3	↓	19	↑↑	21	↑↑	0	None
07NB001	-9	↓	0	None	11	↑↑	-0.1	↓
FortProvidence	-	-	-	-	16	-	-	-
10GC001	-8	↓↓	-4	↓	6	↑↑	0.1	↑
10KA001	-	-	-	-	-	-	-	-
10LC014	-	-	-	-	-	-	-	-
10MC002	-	-	-	-	-	-	-	-
08BB005	0	None	4	↑	7	↑	1.8	↑↑
08CE001	0	None	-4	↓	-3	↓	0.3	↑
08CG001	-4	↓	-5	↓↓	0	None	0.9	↑
08DB001	-7	↓	-5	↓	0	None	0.3	↑
08EF001	-12	↓↓	-7	↓↓	2	↑	0.1	↑
08FF001	-1	↓	10	↑	6	↑	-1.1	↓
08FA002	-4	↓	1	↑	9	↑	-0.8	↓
08FB006	-	-	-	-	-	-	-	-
08FB007	-	-	-	-	-	-	-	-
08FC003	-	-	-	-	-	-	-	-
08GD004	-	-	-	-	-	-	-	-
08MF005	-14	↓↓	-4	↓	10	↑↑	-0.3	↓
08NM085	37	↑	23	↑	0	None	0	None
08NE058	-4	↓	14	↑↑	15	↑	-0.8	↓↓
05DF001	-11	↓	6	↑	18	↑	-0.2	↓
05GG001	-14	↓↓	4	↑	20	↑↑	-0.1	↓
05AJ001	41	↑↑	28	↑↑	-8	↓	-0.1	↓
05CK004	12	↑	30	↑↑	-3	↓	0	None
05HG001	-	-	-	-	-	-	-	-
05KJ001	-11	↓↓	7	↑	13	↑	0	None

Table 5.6 (con't). Freshet trends results for the 1966-2010 (45-year) period.

Station	Pulse Date++		Centre of Mass++		Length+		Volume+	
	#	%	#	%	#	%	#	%
# ↓	16	64%	6	24%	4	15%	14	56%
# ↓↓	6	24%	2	8%	0	0%	4	16%
# ↑	5	20%	17	68%	17	65%	7	28%
# ↑↑	1	4%	4	16%	5	19%	1	4%
# No trend	4	16%	2	8%	5	19%	4	16%
# Missing	12	48%	12	48%	11	42%	12	48%

+ Variable is field significant at 10% significance level.
++ Variable is field significant at 5% significance level.
↑ or ↓: Non-significant trend
↑↑ or ↓↓: Trend significant at 10% significance level
↑↑ or ↓↓: Trend significant at 5% significance level

5.2 Principal Component Analysis

The results of trend analysis revealed that spatial coherence in the direction of trend was relatively strong, particularly during the cold seasons (winter, fall, and the November-April cold season), but also on an annual basis. A pairwise scatterplot matrix is a useful tool for examining runoff covariance in order to determine spatial coherence. Figure 5.14 shows a scatterplot correlation matrix for 1976-2010 annual runoff anomalies for a subset of eight CROCWR basins. (Since the entire CROCWR matrix is comprised of 25 rows and 25 columns, only eight are shown for space considerations.) Runoff anomaly values for each pair of basins compared are shown on the x- and y-axes. The complete (numerical) 35-year annual runoff correlation matrix showing Pearson correlation coefficient values for each pair of watersheds is located in Appendix C, Table C.1. The matrix highlights some strong covariance among certain basins. For example, Lower Athabasca River runoff covaries strongly with Smoky River and Lower Peace River runoff, but not with Upper Liard River runoff.

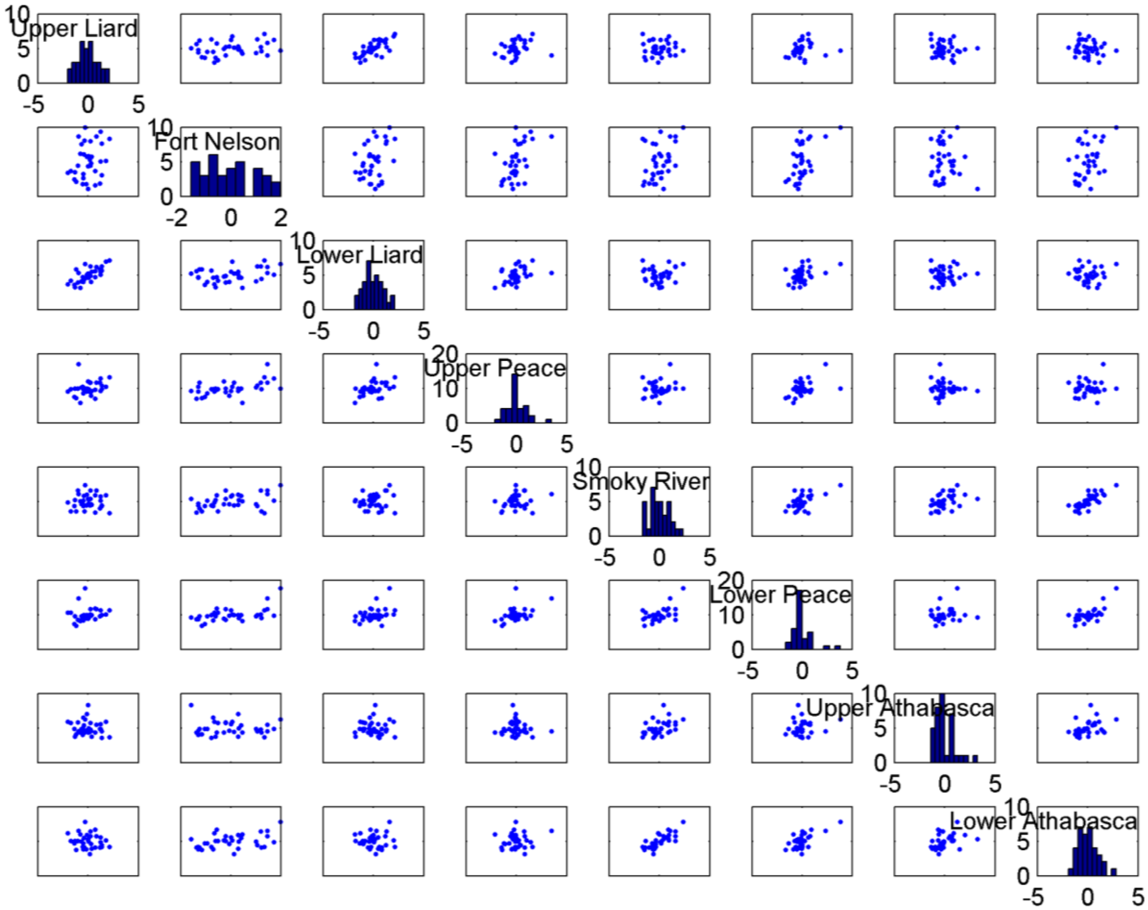


Figure 5.14. Scatterplot correlation matrix of 1976-2010 annual runoff for eight CROCWR basins.

Figure 5.14 also shows that there tends to be spatial coherence with respect to the basins that exhibited stronger correlations. For example, the Smoky River and Lower Athabasca basins are located adjacent to one another, but are not sourced from the same headwater rivers. This suggests that there is some external factor (e.g., climate or geology) that controls runoff in both of these basins.

PCA was performed on both runoff and streamflow timing variables. Due to the extent of missing data for the longer time period, PCA was performed only for the 35-year period. Deciding the number of components to retain presented a challenge, as none of the criteria (described in *Section 4.2.2*) provided an obvious solution. For the runoff variables, between six and eight PCs yielded values greater than one, however graphical interpretation of these modes revealed no clear pattern. For the freshet variables, up to ten eigenvalues exhibited magnitudes of one or greater, therefore the Kaiser-Guttman criterion was not used. Morrison's (2005) advice of retaining a maximum five PCs that account for a minimum 80% of the variance was violated in all cases, therefore this method was also not applied. For all variables, the scree plot suggested retaining between the first one and four PCs. The solution was therefore tested using both three and four PCs, as it was felt that retaining only one or two components would not adequately capture all important regions of hydrologic variability within the CROCWR study area. Varimax rotation of the first three and four modes was performed and mapped, and it was determined that three factors captured the best regionalization of both runoff and hydrological timing variability. Table 5.7 provides the percentage of variability accounted for by each of the first three PCs for each variable.

Table 5.7. Percent variance explained by the first three rotated PCs for each variable.

Variable	PC1	PC2	PC3	Total
		Runoff		
Annual	19.5%	16.1%	18.2%	**53.8%**
Winter	19.4%	15.1%	12.4%	**46.9%**
Spring	21.0%	13.9%	17.4%	**52.3%**
Summer	20.1%	18.9%	15.6%	**54.6%**
Fall	20.6%	14.6%	11.7%	**46.9%**
Cold Season	22.8%	15.4%	10.8%	**49.0%**
Warm Season	21.3%	14.9%	17.4%	**53.6%**
		Freshet		
Pulse Date	16.5%	11.5%	15.8%	**43.8%**
CM Date	15.8%	16.3%	15.5%	**47.6%**
Length	16.8%	14.2%	11.7%	**42.7%**
Volume	22.4%	19.8%	16.8%	**59.0%**

For visual analysis of runoff PCA, each watershed received a unique factor loading to each PC that covered the entire watershed. For the freshet variables, the factor loadings at each WSC station were interpolated using Inverse Distance Weighting (IDW) for facilitated interpretation. IDW is an exact interpolator used to create a surface where interpolated values are most influenced by the closest points, and less influenced by points farther away (O'Sullivan & Unwin, 2010; Linton *et al.*, 2013). A factor loading was considered significant if its magnitude was greater than or equal to absolute value 0.32. Factor loadings with magnitudes of 0.32 to 0.55

were considered significant but weak; loadings of 0.55 to 0.78 were considered moderate; and loadings of 0.78 to 1.0 were considered strong.

For each variable, the runoff or freshet data matrix was projected onto the PC loadings matrix to obtain time series for each PC. The correlation matrix was specified in the PCA to avoid loadings being focused exclusively in regions of high runoff (Maurer *et al.*, 2004), hence PC factor loadings represent patterns of spatially coherent runoff anomalies (Lins, 1997). MK trend analysis was performed on the PC scores, providing a source of comparison between watershed-scale trend results. Figures 5.15 and 5.16 show factor loading maps and corresponding score plots of the first three rotated PCs for annual runoff and pulse date, respectively. PCA results for winter, spring, summer, fall, cold season, and warm season runoff, as well as CM date, freshet length, and freshet volume are presented in Appendix C, Figures C.1 through C.9.

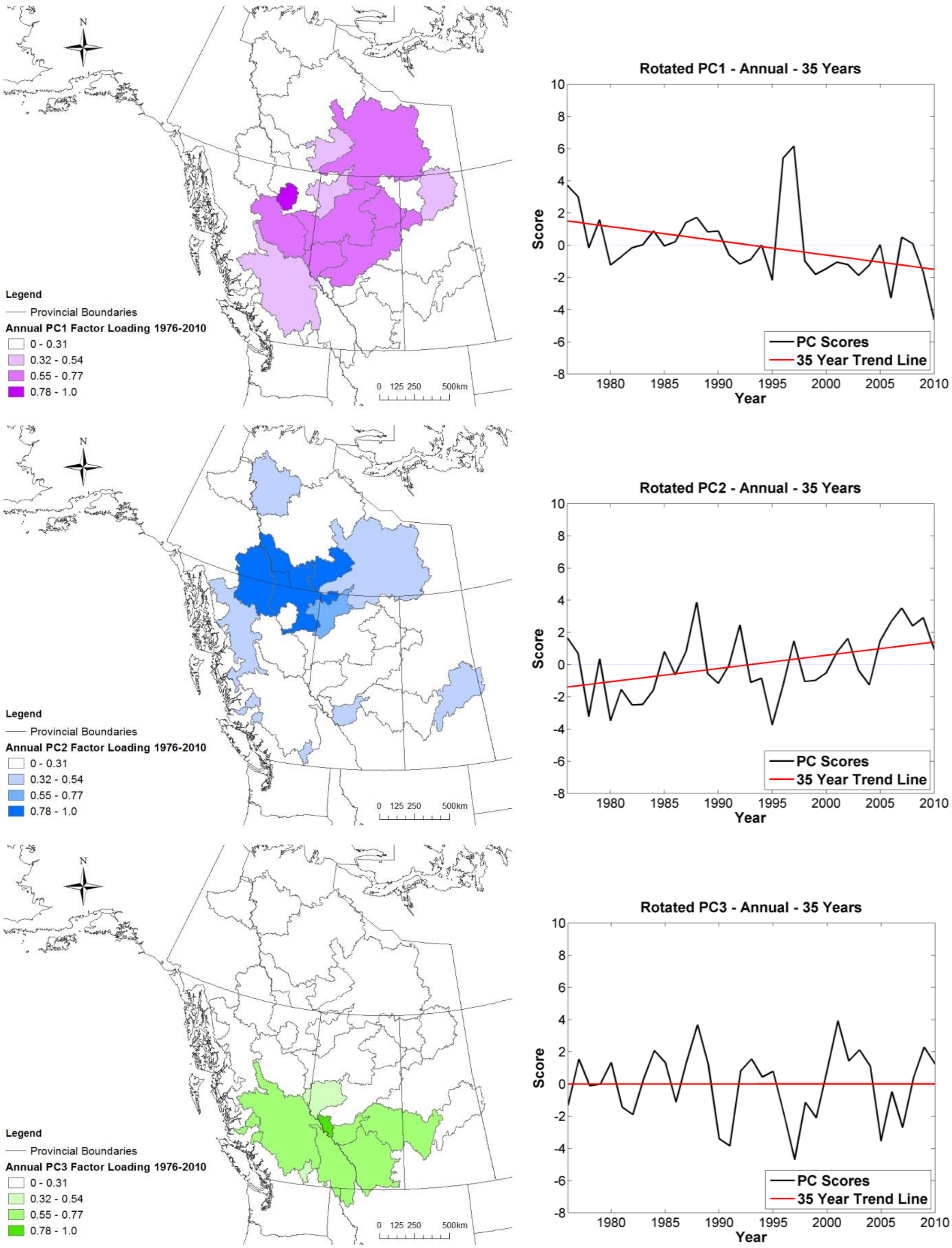


Figure 5.15. Factor loading maps for the first three rotated PCs for annual runoff (left hand side) and corresponding score plots for each PC (right hand side).

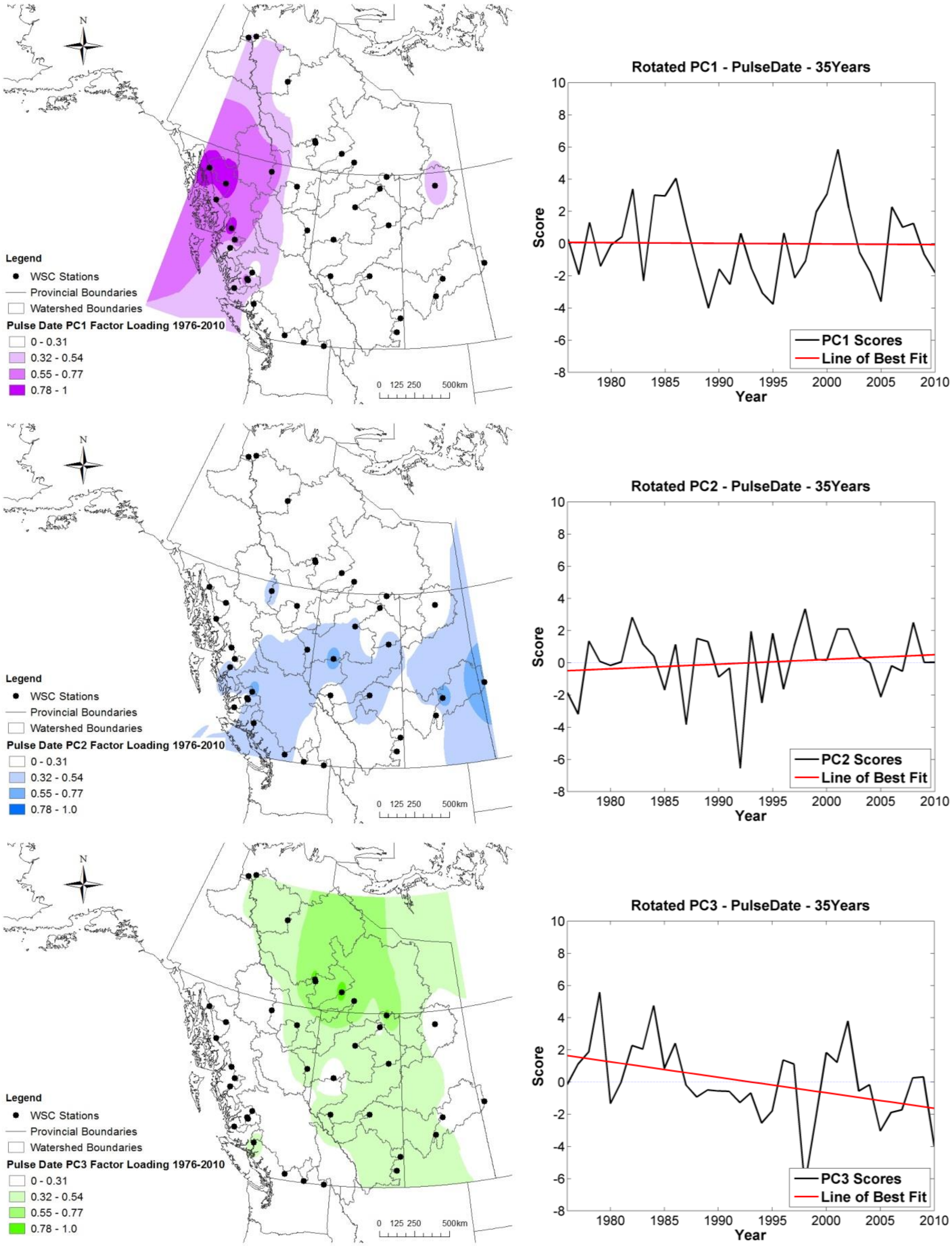


Figure 5.16. Factor loading maps for the first three rotated PCs for pulse date (left hand side) and corresponding score plots for each PC (right hand side).

Certain watersheds/stations did not load significantly to any of the first three rotated PCs, while others loaded significantly to more than one PC. Maps showing the single PC to which each basin or station loaded the highest are given in Figure 5.17 for annual runoff and Figure 5.18 for pulse date. This provides a simple clustering of the basins/stations and helps to distinguish what group a basin/station that loaded significantly to zero or more than one PC is most similar. Winter, spring, summer, fall, cold season, and warm season runoff, as well as CM date, freshet length, and freshet volume results are given in Appendix C, Figures C.10 to C.14.

Rotated PCA of annual runoff revealed three regions of coherent hydrological variability: the high latitudes or "northern region" (NR), the mid-latitudes or "middle region" (MR) and the lower latitudes or "southern region" (SR). These three regions were also prevalent for the summer and warm season PCA, and to a lesser extent the spring. Winter, fall, and cold-season PCA showed somewhat less obvious "regions", however modes were matched as best as possible to reflect the NR, MR, and SR. The following descriptions therefore refer to only the annual, spring, summer, and warm season runoff PCA results.

The NR included mainly watersheds that exhibited increasing trends or tendencies on both an annual and seasonal basis. For the warm season and annually, the Mid-Mackenzie and Peel River basins again differed from their surroundings by not loading significantly to the NR PC. For summer, all watersheds north of 60°N loaded to the NR, whereas for the spring, watersheds concentrated in the Liard River and adjacent basins formed the NR. The MR included mainly watersheds that displayed decreasing runoff trends and tendencies annually and seasonally, including the Peace-Athabasca River system. The MR included the largest number of basins during the spring and warm seasons, extending both northward and southward. The SR generally consisted of the lowest latitude basins, however the region also extended north during spring, encompassing the North Pacific basin.

For the freshet PCA, three regions were prevalent for all variables: the western region (WR), southeastern region (SER), and northern region (FNR). Just as the names imply, PC regions included stations to the west, southeast, and north of the study region. For freshet length, the WR and SER were virtually indistinguishable, as stations loading to these two regions were interspersed within the the same regions; this implies that perhaps retaining three PCs was not the best choice for this variable.

MK trend analysis was performed on the runoff and freshet variable PC scores, providing a source of comparison between watershed-scale and station-specific trend results. Table 5.8 and Table 5.9 provide PC score trend results for runoff and freshet variables, respectively. Trend slopes are given in units of anomaly per year for runoff variables and anomaly/35yrs for freshet variables. The PC number (1 to 3) to which each region corresponded is also given. Runoff trends in units of anomaly/35yrs are provided in Appendix C, Table C.2.

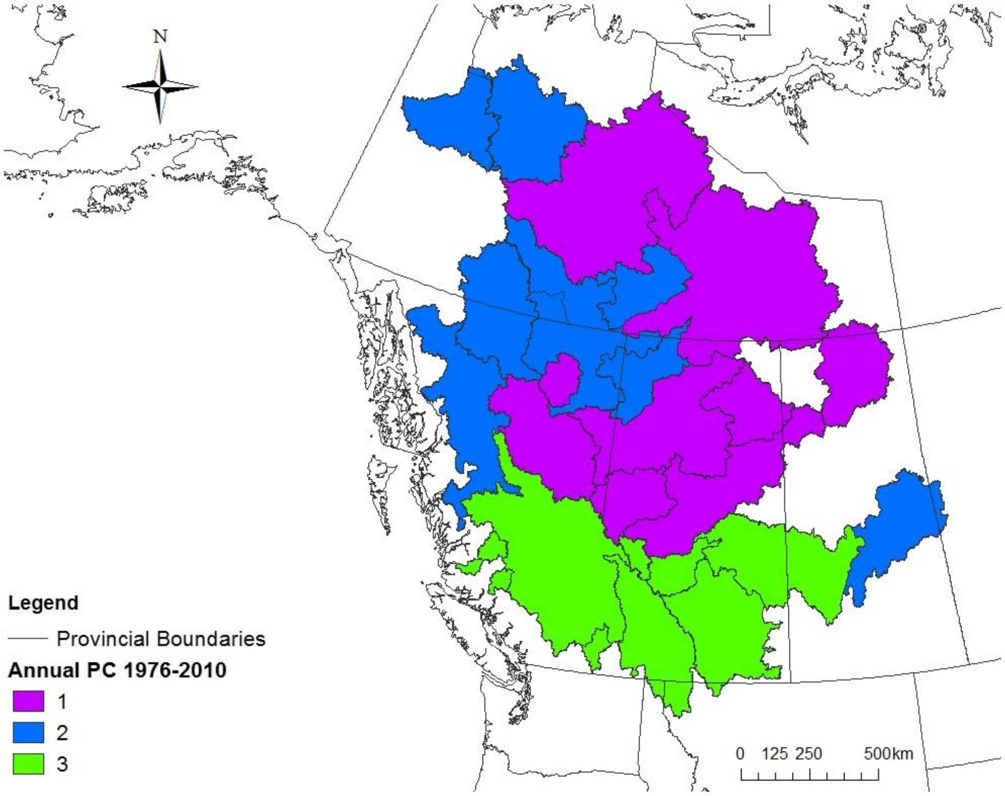


Figure 5.17. Factor loading to which each basin loaded highest in annual runoff PCA.

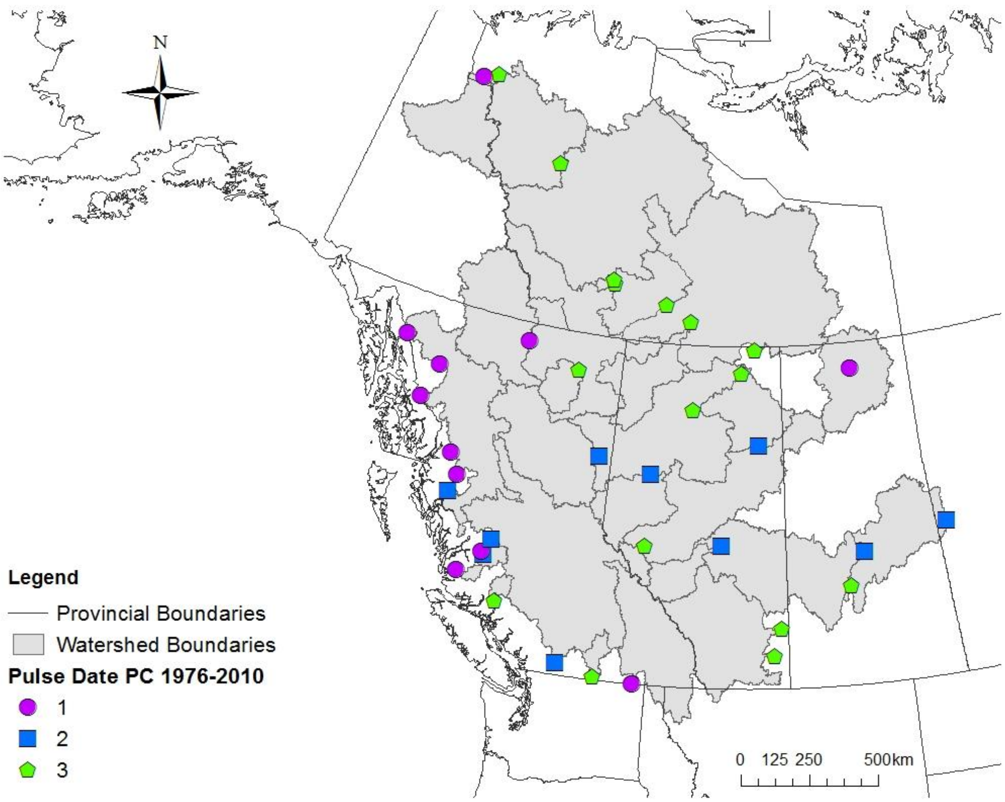


Figure 5.18. Factor loading to which each station loaded highest in pulse date PCA.

Table 5.8. PC region trend results for runoff variables.

		Northern Region	Middle Region	Southern Region
Annual (mm/yr)	**Slope**	0.1	-0.1	0
	Trend	↑↑	↓↓	None
	PC	2	1	3
Winter (mm/yr)	**Slope**	0	0.1	0
	Trend	None	↑↑	None
	PC	3	2	1
Spring (mm/yr)	**Slope**	0	-0.1	0
	Trend	None	↓↓	None
	PC	2	1	3
Summer (mm/yr)	**Slope**	-0.1	-0.1	0
	Trend	↓	↓	None
	PC	2	1	3
Fall (mm/yr)	**Slope**	0.1	-0.1	0.1
	Trend	↑↑	↓↓	↑↑
	PC	3	1	2
Cold Season (mm/yr)	**Slope**	0.1	0.1	0
	Trend	↑	↑↑	None
	PC	3	2	1
Warm Season (mm/yr)	**Slope**	0.1	-0.1	0
	Trend	↑↑	↓↓	None
	PC	2	1	3

Table 5.9. PC region trend results for freshet variables.

		Western Region	Southeastern Region	Northern Region
Pulse Date (days/35yrs)	**Slope**	0	1	-3
	Trend	None	↑	↓↓
	PC	1	2	3
CM Date (days/35yrs)	**Slope**	-1	0	-1
	Trend	↓	None	↓
	PC	1	2	3
Length (days/35yrs)	**Slope**	-2	-3	2
	Trend	↓	↓↓	↑↑
	PC	1	2	3
Volume (mm/35yrs)	**Slope**	1.8	0.8	3.1
	Trend	↑	↑	↑↑
	PC	1	2	3

↑ or ↓: Non-significant trend

↑↑ or ↓↓: Trend significant at 10% significance level

↑↑ or ↓↓: Trend significant at 10% significance level

Regional trend analysis of runoff PC scores confirmed observations that the higher latitudes experienced significantly increased runoff, while the mid-latitudes have undergone reduced runoff, at least annually and during the warm season. For the summer, NR trend results indicated a slight decreasing trend, however the Lower Athabasca and Upper and Lower South Saskatchewan River basins also loaded to the summer NR PC, each of which individually experienced decreased (or no change in) runoff during summer. For spring, the NR increasing trend was very slight, such that on an annual scale, it was virtually undetectable (0 mm/yr). The annual and warm-season increases in high-latitude runoff and related drying of the mid-latitude subtropics reflect documented north-south contrasts in precipitation patterns (e.g., Serreze *et al.*, 2000; Zhang *et al.*, 2007; Min *et al.*, 2008).

For the winter and cold season, only MR displayed a significant trend in the positive direction. It is important to reiterate, however, that the winter and cold season MR differed from the annual and warm season MR. The cold-season MR encompassed a number of northern basins, including the Liard and North Pacific River basins, therefore it was not unexpected to see an increasing trend here. The winter MR, on the other hand, consisted of basins in the Athabasca-Lower Saskatchewan Rivers, the majority of which experienced no trend or slightly decreased runoff during winter, therefore it was surprising to see an increasing trend in MR PC.

For the freshet, the WR experienced no overall change in pulse date, but an advance in the timing of the end of the freshet, resulting in a decreased length of the spring freshet. The volume during this shortened freshet period, however, increased over the 35-year period. In the SER, the pulse date shifted toward a later timing, while the CM date experienced no overall change, again resulting in a shorter spring freshet, but with no change in volume. In the FNR, both the onset and end of the spring freshet advanced in timing, where the magnitude of the pulse date trend was greater than that of the CM date; the length of the spring freshet in the FNR was therefore lengthened, resulting in an associated increase in the volume of the spring freshet in the North. This result aligns well with the results of both watershed-scale and runoff PCA trend analysis, which showed increases in warm-season runoff in the North.

5.3 Climatic Links to Hydrology

Investigation into the influences of near-surface climatic and large-scale atmospheric drivers was performed for the 35-year analysis period of 1976-2010 only.

5.3.1 Correlation with Surface Climate

Prior to examining the influence of surface climate variables on runoff and streamflow timing in the CROCWR region, trend analysis was performed on air temperature and precipitation variables. Table 5.10 and Table 5.11 show trend results for mean temperature and total precipitation, respectively. Annual trend maps are shown in Figure 5.19. Results for maximum and minimum temperatures are located in Appendix D, Tables D.1 and D.2.

Table 5.10. Mean temperature trend results for the 1976-2010 (35-year) period.

Watershed	Annual[+] (°C/35yrs)		Winter (°C/35yrs)		Spring[+] (°C/35yrs)		Summer[++] (°C/35yrs)		Fall (°C/35yrs)		Cold Season (°C/35yrs)		Warm Season[+] (°C/35yrs)	
	Slope	**Trend**	**Slope**	**Trend**	**Slope**	**Trend**	**Slope**	**Trend**	**Slope**	**Trend**	**Slope**	**Trend**	**Slope**	**Trend**
Upper Liard	0.6	↑	-1.6	↓	1.1	↑	0.9	↑↑	-0.8	↓	0.2	↑	0.7	↑
Fort Nelson	0.6	↑	-0.4	↓	-0.1	↓	0.5	↑	0.1	↑	0.9	↑	0	None
Lower Liard	0.5	↑	-0.2	↓	0.6	↑	0.6	↑	0.2	↑	0.8	↑	0.3	↑
Upper Peace	0.4	↑	-0.9	↓	0.2	↑	0.4	↑	-0.3	↓	0.3	↑	0.1	↑
Smoky River	-0.3	↓	-0.4	↓	-1.1	↓	0.2	↑	-0.7	↓	-0.1	↓	-0.6	↓
Lower Peace	0.8	↑	0.4	↑	-0.5	↓	0.5	↑	0.8	↑	0.9	↑	0	None
Upper Athabasca	0.8	↑	2	↑	-0.4	↓	1.1	↑↑	0.5	↑	1.6	↑	0.1	↑
Lower Athabasca	0.3	↑	0.7	↑	-0.8	↓	0.7	↑	0.7	↑	1.1	↑	-0.1	↓
East Lake Athabasca	0.9	↑↑	1.7	↑	0	None	1	↑↑	2.3	↑	1.6	↑	0.5	↑
West Lake Athabasca	1	↑	1.8	↑	-0.5	↓	0.8	↑↑	2.4	↑↑	2	↑	0.2	↑
Hay	0.8	↑	0.5	↑	-0.3	↓	0.2	↑	2	↑	1.6	↑	-0.1	↓
Great Slave	1.7	↑↑	2	↑	0.8	↑	1.1	↑↑	3.4	↑↑	2.8	↑↑	0.6	↑
Upper Mackenzie	1.2	↑	1.3	↑	0.4	↑	0.4	↑	2.5	↑↑	2.3	↑	0.2	↑
Mid Mackenzie	1.1	↑	0.8	↑	0.9	↑	0.2	↑	1.7	↑	1.8	↑	0.2	↑
Lower Mackenzie	1.3	↑↑	0.3	↑	2.9	↑↑	0.5	↑	1.6	↑	2	↑	0.8	↑↑
Peel	1.4	↑	0.5	↑	2.6	↑↑	0	None	1.8	↑	2.2	↑	0.9	↑↑
North Pacific	0.5	↑	-0.3	↓	0.7	↑	0.5	↑	-0.2	↓	0.2	↑	0.3	↑
South Pacific	0.9	↑	0.3	↑	0.8	↑	1.2	↑↑	-0.2	↓	0.7	↑	0.9	↑↑
Fraser	0.7	↑	1.2	↑	0.4	↑	1	↑↑	0	None	0.8	↑	0.5	↑
Okanagan	1.4	↑↑	1.3	↑	1	↑	2.3	↑↑	0.8	↑	1.2	↑↑	1.6	↑↑
Columbia	1.1	↑↑	1.1	↑	0.2	↑	1.6	↑↑	0.6	↑	1.2	↑↑	0.9	↑↑
Upper North Saskatchewan	0.8	↑	2.4	↑	-0.6	↓	0.8	↑↑	0.9	↑	1.8	↑↑	-0.1	↓
Lower North Saskatchewan	0.2	↑	0.5	↑	-1.1	↓↓	0.8	↑↑	0.7	↑	0.7	↑	-0.2	↓
Upper South Saskatchewan	0.3	↑	1.5	↑	-1.1	↓	0.6	↑	0.6	↑	1.2	↑	-0.1	↓
Lower South Saskatchewan	0.5	↑	0.8	↑	-1.2	↓	1	↑↑	1.9	↑	1.3	↑	0.2	↑

Table 5.10(con't). Mean temperature trend results for the 1976-2010 (35-year) period.

Station	Annual+		Winter		Spring+		Summer++		Fall		Cold Season		Warm Season+	
	#	%	#	%	#	%	#	%	#	%	#	%	#	%
# ↓	1	**4%**	6	**24%**	10	**40%**	0	**0%**	5	**20%**	1	**4%**	6	**24%**
# **↓↓**	0	**0%**	0	**0%**	1	**4%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**
# ↑	19	**76%**	19	**76%**	11	**44%**	12	**48%**	16	**64%**	20	**80%**	12	**48%**
# **↑↑**	5	**20%**	0	**0%**	2	**8%**	12	**48%**	3	**12%**	4	**16%**	5	**20%**
# No trend	0	**0%**	0	**0%**	1	**4%**	1	**4%**	1	**4%**	0	**0%**	2	**8%**
# Missing	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**

+ Variable is field significant at 10% significance level.

++ Variable is field significant at 5% significance level.

↑ or ↓: Non-significant trend

↑↑ or **↓↓**: Trend significant at 10% significance level

↑↑ or **↓↓**: Trend significant at 10% significance level

Table 5.11. Preciptation trend results for the 1976-2010 (35-year) period.

Watershed	Annual[++] (mm/yr)		Winter[++] (mm/yr)		Spring[+] (mm/yr)		Summer[++] (mm/yr)		Fall[++] (mm/yr)		Cold Season[++] (mm/yr)		Warm Season (mm/yr)	
	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend
Upper Liard	-1.4	↓	0.3	↑	0.1	↑	-0.3	↓	-1.4	↓↓	-0.8	↓	-0.3	↓
Fort Nelson	1.2	↑	0.9	↑↑	0.4	↑	0.5	↑	-0.5	↓	0.0	None	1.2	↑
Lower Liard	-0.2	↓	0.5	↑↑	0.1	↑	-0.3	↓	-0.4	↓	-0.3	↓	-0.2	↓
Upper Peace	-1.6	↓	0.2	↑	-0.3	↓	-0.8	↓	-1.1	↓	-1.3	↓	-0.8	↓
Smoky River	-4.5	↓↓	-0.3	↓	-1.3	↓↓	-2.4	↓↓	-0.9	↓	-1.4	↓↓	-3.6	↓↓
Lower Peace	-2.1	↓	0.3	↑	-0.8	↓	-0.9	↓↓	-0.8	↓	-0.3	↓	-1.8	↓
Upper Athabasca	-6.0	↓↓	-1.8	↓↓	0.6	↑	-1.5	↓↓	-2.5	↓↓	-4.3	↓↓	-1.8	↓↓
Lower Athabasca	-3.4	↓↓	-0.3	↓	-0.4	↓	-1.5	↓↓	-1.0	↓↓	-1.1	↓↓	-2.4	↓↓
East Lake Athabasca	-1.2	↓	0.1	↑	-0.5	↓	0.1	↑	-1.0	↓	-0.8	↓	-0.4	↓
West Lake Athabasca	-1.8	↓↓	-0.4	↓	-0.3	↓	0.2	↑	-1.5	↓↓	-1.7	↓↓	-0.2	↓
Hay	-0.3	↓	0.9	↑↑	-0.7	↓	0.2	↑	-0.5	↓	0.3	↑	-0.3	↓
Great Slave	-1.2	↓	-0.1	↓	-0.5	↓↓	0.3	↑	-1.0	↓↓	-0.8	↓↓	-0.2	↓
Upper Mackenzie	-0.2	↓	-0.1	↓	-0.2	↓	1.0	↑	-0.5	↓	-1.0	↓↓	1.0	↑
Mid Mackenzie	-1.3	↓↓	-0.1	↓	-0.4	↓↓	0.1	↑	-0.8	↓↓	-0.9	↓↓	-0.3	↓
Lower Mackenzie	-1.1	↓	0.0	None	-0.6	↓↓	0.0	None	-0.7	↓↓	-0.1	↓	-0.8	↓
Peel	-2.6	↓↓	-0.4	↓	-0.8	↓↓	-0.3	↓	-1.1	↓	-1.2	↓↓	-1.3	↓↓
North Pacific	5.9	↑	3.0	↑↑	0.9	↑↑	2.1	↑↑	-1.0	↓	2.8	↑	3.3	↑↑
South Pacific	-1.2	↓	0.5	↑	0.1	↑	-0.6	↓	-0.7	↓	-0.9	↓	0.4	↑
Fraser	-2.2	↓	0.1	↑	-0.3	↓	-1.0	↓↓	-1.1	↓	-1.5	↓	-1.1	↓
Okanagan	-4.0	↓↓	-0.2	↓	0.0	None	-1.8	↓↓	-1.2	↓↓	-2.4	↓↓	-1.6	↓
Columbia	-7.0	↓↓	-1.3	↓	0.2	↑	-1.8	↓↓	-4.4	↓↓	-6.0	↓↓	-2.1	↓↓
Upper North Saskatchewan	-3.7	↓↓	-0.8	↓	0.3	↑	-1.4	↓↓	-1.7	↓↓	-2.3	↓↓	-2.5	↓↓
Lower North Saskatchewan	-1.0	↓	-0.3	↓	-0.1	↓	-0.4	↓	-0.6	↓↓	-0.8	↓↓	-0.7	↓
Upper South Saskatchewan	-1.1	↓	-0.4	↓↓	1.6	↑↑	-1.0	↓	-0.9	↓↓	-1.4	↓↓	0.0	None
Lower South Saskatchewan	-0.9	↓	0.0	None	-0.1	↓	0.2	↑	-0.9	↓↓	-0.5	↓	-0.3	↓

Table 5.11(con't). Preciptation trend results for the 1976-2010 (35-year) period.

Station	Annual[++]		Winter[++]		Spring[+]		Summer[++]		Fall[++]		Cold Season[++]		Warm Season	
	#	%	#	%	#	%	#	%	#	%	#	%	#	%
# ↓	14	**56%**	11	**44%**	10	**40%**	7	**28%**	12	**48%**	9	**36%**	14	**56%**
# **↓↓**	9	**36%**	2	**8%**	5	**20%**	8	**32%**	13	**52%**	13	**52%**	6	**24%**
# ↑	2	**8%**	6	**24%**	7	**28%**	8	**32%**	0	**0%**	2	**8%**	3	**12%**
# **↑↑**	0	**0%**	4	**16%**	2	**8%**	1	**4%**	0	**0%**	0	**0%**	1	**4%**
# No trend	0	**0%**	2	**8%**	1	**4%**	1	**4%**	0	**0%**	1	**4%**	1	**4%**
# Missing	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**

[+] Variable is field significant at 10% significance level.

[++] Variable is field significant at 5% significance level.

↑ or ↓: Non-significant trend

↑↑ or **↓↓**: Trend significant at 10% significance level

↑↑ or ↓↓: Trend significant at 10% significance level

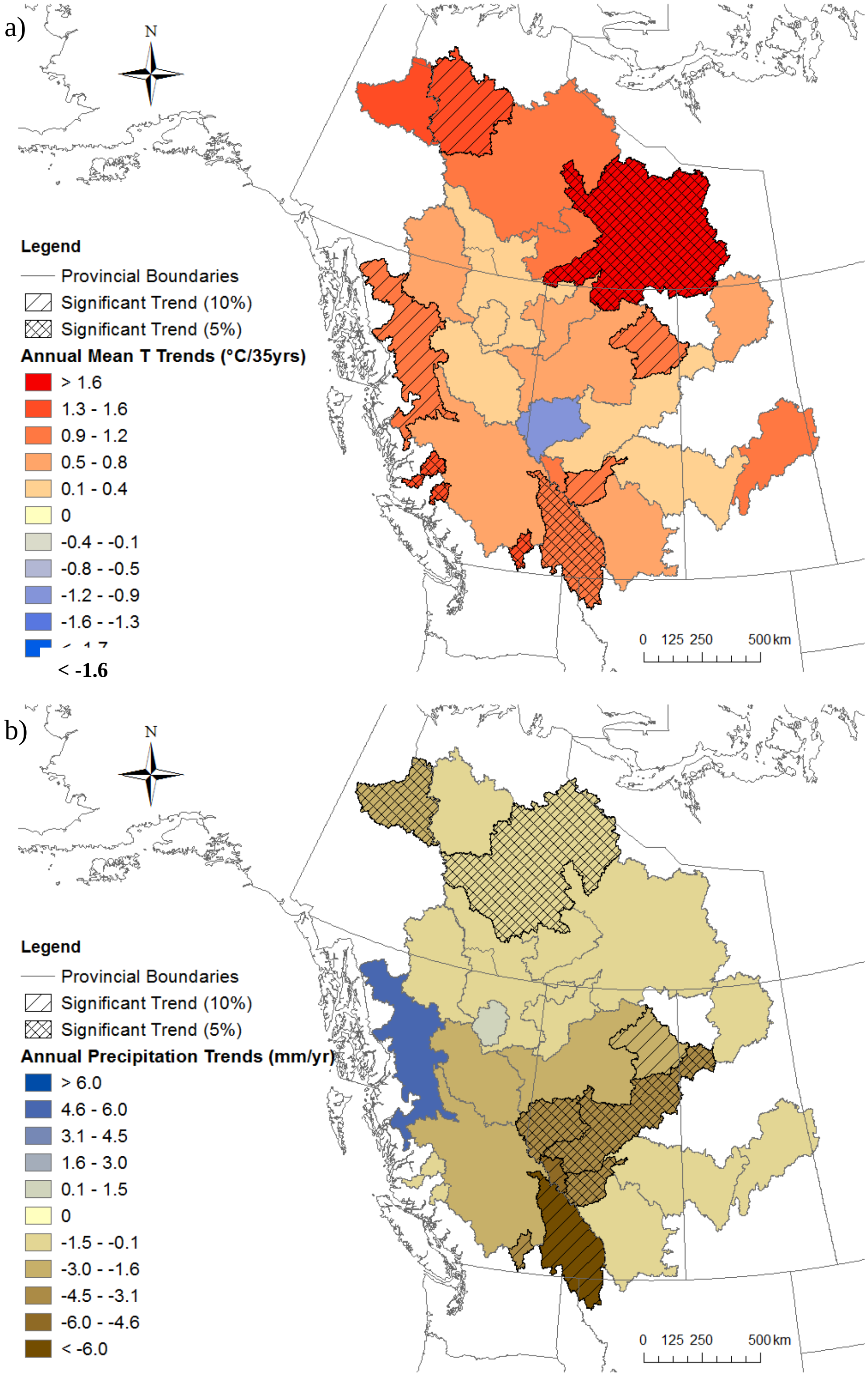


Figure 5.19. Map of a) mean annual temperature and b) total annual precipitation trend slopes for the 1976-2010 analysis period. Basins exhibiting significant trends are shown as hatched.

For all measures of temperature, the vast majority of watersheds exhibited no trend on a per year basis for the 35-year period; thus, temperature results are presented in units of °C per 35-year analysis period. Temperature was shown to have increased in the majority of basins both annually and during each season, except for spring mean and minimum temperature and warm-season minimum temperature, when decreasing trends were detected in approximately half of the basins. Decreasing temperatures during these seasons were mainly concentrated in the Peace-Athabasca-Saskatchewan River basins, and could help explain the unusal patterns noted in spring runoff (with respect to the other seasons), especially in the more southern and middle latitudes.

For precipitation, the majority of basins exhibited decreasing trends, particularly on an annual basis and during the fall and cold season. Winter, spring, summer, and warm season exhibited more increases in precipitation, though with the exception of the Pacific watersheds, basins that did experience increased precipitation were generally not the same for each of these four time variables. The greatest number of significant increasing trends in precipitation occurred during the winter; these basins were concentration in the Liard-North Pacific region.

Each runoff variable was correlated with maximum, mean, and minimum temperature and total precipitation data for the corresponding year or season (lag 0), as well as for up to four seasonal time lags. For cold- and warm-season runoff, the influence of temperature and precipitation during the preceding warm and cold seasons, as well as during the cold and warm seasons one-year prior, were analyzed. For winter, spring, summer, and fall runoff, the influences of climate up to four seasons before were studied.

The number of basins that showed a significant ($p < 0.1$) Pearson correlation between runoff and each climate metric is given in Table 5.12, as is the percent out of 25 basins total. Results are shown if a significant relationship existed in six or more basins. Average Pearson R values (R_{avg}) for the basins that experienced significant correlations are also provided, as is the relationship (positive or negative) between the runoff and climate variables (i.e., a negative relationship between maximum temperature and annual runoff would imply that the majority of basins that exhibited a significant correlation between maximum temperature and annual runoff experienced decreased runoff with increased temperature and vice versa). Bolded rows indicate relationships in which more than half of the basins exhibited a significant correlation.

Table 5.12. Significant correlations between runoff and climate variables.

Runoff Variable	Climate Variable	Type	Season	Lag	# Basins	% Basins	Relationship+	R_{avg}*
				Runoff				
Annual	Temperature	Max	Ann	0	9	36%	Negative	-0.35*
	Temperature	Mean	Ann	0	8	32%	Negative	-0.35*
	Temperature	Min	Ann	0	8	32%	Mixed	-0.34,0.36
	Precipitation	**Total**	**Ann**	**0**	**21**	**84%**	**Positive**	**0.6**
	Precipitation	Total	Ann	1	9	36%	Positive	0.45
Winter	Temperature	Max	Winter	0	7	28%	Positive	0.45*
	Temperature	Mean	Winter	0	6	24%	Positive	0.48*
	Temperature	Min	Winter	0	6	24%	Positive	0.48*
	Precipitation	Total	Winter	0	6	24%	Positive	0.39*
	Precipitation	Total	Fall	1	9	36%	Positive	0.44*
	Precipitation	**Total**	**Summer**	**2**	**14**	**56%**	**Positive**	**0.45**
	Precipitation	Total	Spring	3	9	36%	Positive	0.38*
Spring	Temperature	Max	Spring	0	11	44%	Negative	-0.4*
	Temperature	Mean	Spring	0	8	32%	Negative	-0.38*
	Temperature	Min	Spring	0	6	24%	Mixed	-0.35,0.34
	Precipitation	**Total**	**Spring**	**0**	**15**	**60%**	**Positive**	**0.49***
	Precipitation	Total	Winter	1	10	40%	Positive	0.47
	Precipitation	**Total**	**Fall**	**2**	**16**	**64%**	**Positive**	**0.44***
	Precipitation	Total	Summer	3	8	32%	Positive	0.38*
Summer	Temperature	Max	Summer	0	8	32%	Negative	-0.34
	Precipitation	**Total**	**Summer**	**0**	**19**	**76%**	**Positive**	**0.55**
	Temperature	Max	Spring	1	8	32%	Negative	-0.49
	Temperature	Mean	Spring	1	8	32%	Negative	-0.45
	Precipitation	**Total**	**Spring**	**1**	**14**	**56%**	**Positive**	**0.44**
	Precipitation	Total	Winter	2	8	32%	Positive	0.43
	Precipitation	Total	Fall	3	10	40%	Positive	0.43*
	Temperature	Max	Summer	4	6	24%	Negative	-0.34*
	Temperature	Mean	Summer	4	7	28%	Negative	-0.35
Fall	Precipitation	Total	Fall	0	7	28%	Positive	0.37*
	Temperature	Max	Summer	1	8	32%	Negative	-0.44
	Precipitation	**Total**	**Summer**	**1**	**20**	**80%**	**Positive**	**0.55**
	Precipitation	Total	Spring	2	6	24%	Positive	0.4
	Temperature	Max	Fall	4	8	32%	Negative	-0.33
	Temperature	Mean	Fall	4	6	24%	Negative	-0.36
	Temperature	Min	Fall	4	6	24%	Negative	-0.36
Cold Season	Temperature	Max	Cold	0	9	36%	Positive	0.43*
	Temperature	Mean	Cold	0	9	36%	Positive	0.45*
	Temperature	Min	Cold	0	10	40%	Positive	0.43*
	Precipitation	Total	Cold	0	10	40%	Positive	0.43*

Table 5.12(con't). Significant correlations between runoff and climate variables.

	Variable	Type	Season	Lag	# Basins	% Basins	Relationship+	R_{avg}*
				Runoff				
Cold	Temperature	Max	Warm	1	6	24%	Negative	-0.42
Season	**Precipitation**	**Total**	**Warm**	**1**	**18**	**72%**	**Positive**	**0.51**
Warm	Temperature	Max	Warm	0	12	48%	Negative	-0.4
Season	**Precipitation**	**Total**	**Warm**	**0**	**18**	**72%**	**Positive**	**0.57**
	Temperature	Max	Cold	1	8	32%	Negative	-0.42*
	Temperature	Mean	Cold	1	8	32%	Negative	-0.42*
	Temperature	Min	Cold	1	7	28%	Negative	-0.45*
	Precipitation	**Total**	**Cold**	**1**	**15**	**60%**	**Positive**	**0.55***
	Precipitation	Total	Warm	2	6	24%	Positive	0.45

+ A "Mixed" relationship indicates that more than two basins exhibited a correlation R value of opposite magnitude from the other basins, thus both positive and negative average R values are given

* When two or less basins exhibited a correlation R value of opposite magnitude than the other basins, these opposite magnitude values were excluded from the R_{avg} calculation

Table 5.12 shows that precipitation, particularly during the summer and warm season, is a strong driver of runoff in western Canada. Runoff during all seasons (with the exception of spring) was significantly positively correlated to summer (or warm season, in the case of the cold and warm season runoff) precipitation in more than 50% of the watersheds. Spring runoff was significantly correlated to both spring precipitation as well as fall precipitation of the previous year. Annually, runoff was significantly influenced by precipitation of the same year.

Although precipitation was shown to be a much stronger driver of runoff than air temperature, as gauged by less than half of the basins showing significant relationships between any measure of temperature and runoff annually or during any season, the influence of temperature remains critical. Temperature was negatively correlated to runoff annually and during all seasons except winter and cold season, implying that the overall warming trend in most basins has caused reduced runoff during the warmer months; this is particularly true among the mid-latitude basins. During the winter and cold season, warming temperatures were associated with increased runoff (and vice versa), which helps to explain the increase in winter and cold reason runoff in the north. This is likely due to the fact that increases in temperature have caused earlier spring melt (moving into the cold season), at least in the most northern watersheds, as well as later autumn freeze-up, meaning more open-water flows during the cold season.

The seasons (lags) during which temperature and precipitation exerted the greatest influence (in terms of number of basins affected) on each runoff variable were further examined. Of the three temperature measures, maximum temperature appeared to have the greatest effect on runoff, therefore maximum temperature was selected. Tables Table 5.13 and Table 5.14 provide R values for each basin that exhibited a significant correlation between runoff and maximum temperature or precipitation, respectively, for the specified season and runoff variable. Figure 5.20 provides annual runoff correlation maps. Correlation maps for the remaining runoff variables are located in Appendix D, Figures D.1 through D.7.

Table 5.13. Runoff vs. maximum temperature for lag with highest number of basins exhibiting significant correlations.

Runoff Variable	Ann	Winter	Spring	Summer	Fall	Cold Season	Warm Season
Max Temp Variable (Lag)	**Ann (0)**	**Winter (0)**	**Spring (0)**	**Summer (0)**	**Summer (1)**	**Cold Season (0)**	**Warm Season (0)**
# of Basins	**9**	**7**	**11**	**8**	**8**	**9**	**12**
Upper Liard				-0.39	-0.37	0.31	-0.40
Fort Nelson					-0.35		-0.30
Lower Liard						0.31	-0.35
Upper Peace	-0.30		-0.48	-0.40			-0.61
Smoky River	-0.45	0.46	-0.36		-0.66		-0.41
Lower Peace	-0.36		-0.38				-0.33
Upper Athabasca	-0.39	0.50	(0.35)			0.43	
Lower Athabasca			-0.38	-0.29	-0.43		-0.31
East Lake Athabasca		(-0.33)				(-0.34)	
West Lake Athabasca	-0.32						
Hay						0.53	
Great Slave							
Upper Mackenzie							
Mid Mackenzie			-0.41	-0.30			-0.30
Lower Mackenzie	(0.34)						
Peel							
North Pacific		0.34		-0.35		0.55	-0.45
South Pacific		0.53	(0.47)			0.50	
Fraser	-0.31	0.55		-0.38	-0.36	0.41	-0.50
Okanagan	-0.35		-0.49	-0.29	-0.47		-0.47
Columbia			-0.40				
Upper North Saskatchewan		0.30			-0.37		
Lower North Saskatchewan	-0.30						
Upper South Saskatchewan			-0.42	-0.30	-0.52		
Lower South Saskatchewan			-0.29			0.40	-0.36
R_{avg}*	-0.35	0.45	-0.40	-0.34	-0.44	0.43	-0.40

*Excluding values in brackets which are opposite in magnitude to remaining values in each variable.

Table 5.14. Runoff vs. precipitation for lag with highest number of basins exhibiting significant correlations.

Runoff Variable	**Ann**	**Winter**	**Spring**		**Summer**	**Fall**	**Cold Season**	**Warm Season**
Precipitation Variable (Lag)	**Ann (0)**	**Summer (2)**	**Spring (0)**	**Fall (2)**	**Summer (0)**	**Summer (1)**	**Warm Season (1)**	**Warm Season (0)**
# of Basins	**21**	**14**	**15**	**16**	**19**	**20**	**18**	**18**
Upper Liard	0.68		0.58		0.57	0.52	0.43	0.53
Fort Nelson	0.62	0.46	0.55		0.78	0.78	0.58	0.67
Lower Liard	0.68	0.36	0.59		0.76	0.60	0.33	0.73
Upper Peace	0.43			0.49	0.35			0.31
Smoky River	0.79	0.42	0.41	0.41	0.88	0.47	0.35	0.75
Lower Peace	0.64			0.43	0.62	0.53	0.30	0.55
Upper Athabasca	0.71	0.34	0.38	0.38	0.49	0.35	0.56	0.54
Lower Athabasca	0.69	0.64	0.47	0.56	0.74	0.69	0.73	0.71
East Lake Athabasca	0.51			0.37				0.32
West Lake Athabasca	0.49	0.55			0.62	0.70	0.51	0.60
Hay	0.63	0.64	0.43		0.50	0.68	0.68	0.69
Great Slave		0.49	(-0.29)			0.50	0.38	
Upper Mackenzie		0.31			0.51	0.68		0.41
Mid Mackenzie	0.36				0.37		0.45	
Lower Mackenzie				(-0.37)				
Peel			0.49	0.43	0.43	0.32		
North Pacific	0.77	0.33	0.43	0.61	0.34			0.55
South Pacific	0.51					0.39		
Fraser	0.73	0.53		0.48	0.46	0.56	0.60	0.37
Okanagan	0.69	0.41	0.41	0.58	0.48	0.64	0.56	0.55
Columbia	0.64			0.42		0.32	0.49	
Upper North Saskatchewan	0.68	0.40	0.76	0.34	0.63	0.68	0.54	0.75
Lower North Saskatchewan	0.28		0.31	0.43		0.32	0.51	
Upper South Saskatchewan	0.65	0.47	0.69	0.35	0.45	0.62	0.61	0.72
Lower South Saskatchewan	0.45		0.31	0.33	0.41	0.61	0.49	0.49
R_{avg}*	0.60	0.45	0.49	0.44	0.55	0.55	0.51	0.57

*Excluding values in brackets which are opposite in magnitude to remaining R values for each variable.

Note: Both spring and fall precipitation were selected for spring runoff, since fall (lag 2) precipitation significantly correlated with a greater number of basins, but spring (lag 0) R_{avg} value was larger.

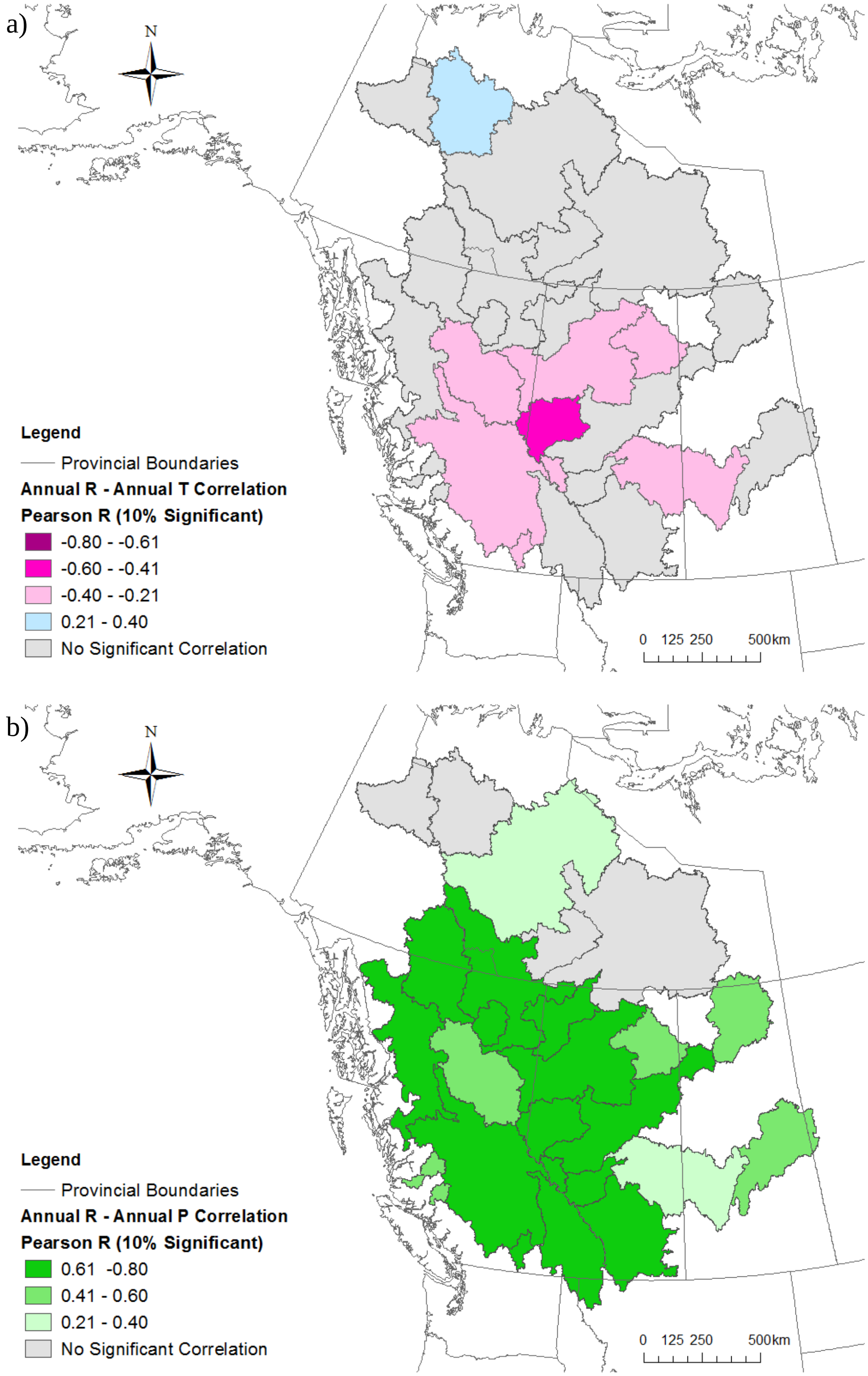


Figure 5.20. Basins that exhibited significant correlations ($p < 0.1$) between annual runoff and a) maximum temperature and b) total precipitation for the same year (lag 0).

Annually, increases in maximum temperature were strongly related to decreased runoff in basins that formed part of the MR PC. For precipitation, it was not surprising to see that the only basins that did not correlate significantly with overall decreasing precipitation trends were those located in the NR which showed generally increasing runoff annually (with the exception of the Mid-Mackenzie River basin, which did correlate with precipitation).

For correlation analysis between climate and freshet timing variables, climate data from the basin upstream of each station were used. Due to the manner in which the watersheds were defined, stations 05HG001 and 07NB001 were not assigned climate data (due to there being no basin upstream of either of these gauges), and were therefore not included in the analysis. Additionally, as discussed in *Section 5.1.3*, station 07FD002 was not included in the analysis. Hence, 34 stations were analyzed.

Since the onset of the spring freshet generally occurs in the early spring, but may also begin earlier in the winter in more southern regions, or later during the summer in some of the more northern rivers, it was of interest to examine the influence of climate for a number of different time frames. Maximum, mean, and minimum temperature and total precipitation for both the winter and spring, as well as for the individual months of January to June, were correlated with each of the freshet variables. Cold-season climate was also examined, as was a new variable encompassing the months of January to April based on the work of Bonsal *et al.* (2006) and hereafter referred to as "Bonsal Spring". Bonsal Spring was shown to well represent relationships between spring river break-up and large-scale atmospheric processes (for use in composite analysis with climate indices, as presented in *Section 5.3.2*) and hence was also applied in this near-surface climate analysis for consistency.

The number of stations that showed a significant ($p < 0.1$) correlation between each freshet variable and each climate metric are given in Table 5.15, as is the percent out of 34 stations. Results are shown if a significant relationship existed at nine or more stations. Average Pearson R values (R_{avg}) for the stations that experienced significant correlations are also provided, as is the relationship (positive or negative) between the freshet and climate variables (i.e., a negative relationship between mean temperature and pulse date would imply that the majority of stations that exhibited a significant correlation between mean temperature and pulse date experienced an earlier onset of the spring freshet with increased temperature and vice versa). Bolded rows indicate relationships at which more than half of the stations exhibited a significant correlation. Bolded R_{avg} values indicate correlations greater than 0.5.

As Table 5.15 shows, only pulse date exhibited a relationship in which over half of the stations were significantly affected by climate. All three measures of temperature during spring, as well as the month of May, were significantly negatively correlated to pulse date. Neither CM date nor freshet length showed any connection to temperature or precipitation in more than half of the stations, however freshet volume did exhibit a significant relationship with cold season precipitation at 46% of the stations. Table 5.16 presents the R values for stations that exhibited a significant correlation between spring mean temperature and pulse date, and cold season precipitation and freshet volume. Maps of each relationship are presented in Figure 5.21.

Table 5.15. Significant correlations between freshet and climate variables.

Variable	Climate Variable	Type	Season	# Stations	% Stations	Relation-ship+	R_{avg}*
			Freshet				
Pulse Date	Temperature	Max	March	10	29%	Mixed	-0.38,0.33
	Temperature	Max	April	16	47%	Negative	**-0.56***
	Temperature	Mean	April	15	44%	Negative	**-0.58**
	Temperature	Min	April	15	44%	Negative	**-0.52**
	Temperature	**Max**	**May**	**17**	**50%**	**Negative**	**-0.49**
	Temperature	**Mean**	**May**	**17**	**50%**	**Negative**	**-0.51**
	Temperature	**Min**	**May**	**17**	**50%**	**Negative**	**-0.5**
	Temperature	Min	June	12	35%	Negative	-0.36
	Temperature	Max	Bonsal Spring	12	35%	Negative	-0.39*
	Temperature	Mean	Bonsal Spring	11	32%	Negative	-0.38*
	Temperature	Min	Bonsal Spring	9	26%	Negative	-0.39*
	Temperature	Max	Winter	9	26%	Mixed	-0.38,0.34
	Temperature	Mean	Winter	9	26%	Mixed	-0.34,0.34
	Temperature	**Max**	**Spring**	**18**	**53%**	**Negative**	**-0.52***
	Temperature	**Mean**	**Spring**	**18**	**53%**	**Negative**	**-0.56***
	Temperature	**Min**	**Spring**	**18**	**53%**	**Negative**	**-0.55***
	Temperature	Max	Cold	13	38%	Negative	-0.36*
	Temperature	Mean	Cold	10	29%	Negative	-0.36*
	Temperature	Min	Cold	9	26%	Negative	-0.35*
	Precipitation	Total	Cold	10	29%	Negative	-0.37*

Table 5.15 (con't). Significant correlations between freshet and climate variables.

Variable	Climate Variable	Type	Season	# Stations	% Stations	Relation-ship+	R_{avg}*
			Freshet				
CM Date	Precipitation	Total	Feb	12	35%	Negative	-0.34*
	Temperature	Max	May	14	41%	Negative	-0.43*
	Temperature	Mean	May	13	38%	Negative	-0.44*
	Temperature	Min	May	12	35%	Negative	-0.43*
	Temperature	Min	Spring	9	26%	Negative	-0.32*
	Precipitation	Total	Cold	11	32%	Negative	-0.35*
Length	Temperature	Max	January	9	26%	Negative	-0.34*
	Temperature	Max	April	16	47%	Positive	0.48
	Temperature	Mean	April	15	44%	Positive	0.47
	Temperature	Min	April	13	38%	Positive	0.43
	Temperature	Min	May	9	26%	Positive	0.45*
	Temperature	Max	Spring	13	38%	Positive	0.43
	Temperature	Mean	Spring	14	41%	Positive	0.42
	Temperature	Min	Spring	12	35%	Positive	0.44
Volume	Precipitation	Total	January	13	38%	Positive	0.35
	Precipitation	Total	March	9	26%	Positive	0.35
	Temperature	Max	June	9	26%	Negative	-0.36*
	Temperature	Mean	June	9	26%	Negative	-0.35*
	Precipitation	Total	Bonsal Spring	13	38%	Positive	0.43*
	Precipitation	Total	Winter	15	44%	Positive	0.42*
	Temperature	Max	Spring	10	29%	Negative	-0.36*
	Precipitation	Total	Spring	9	26%	Positive	0.45*
	Precipitation	Total	Cold	16	47%	Positive	**0.53***

+ A "Mixed" relationship indicates that more than two watersheds/stations exhibited a correlation R value of opposite magnitude from the other watersheds/stations, thus both positive and negative average R values are given

* When two or less watersheds/stations exhibited a correlation R value of opposite magnitude than the other watersheds/stations, these opposite magnitude values were excluded from the R_{avg} calculation

Table 5.16. Spring freshet measure vs. mean temperature and precipitation for seasons with highest percentages of significant correlations.

Freshet Var.	**Pulse Date**	**Volume**	**Freshet Var.**	**Pulse Date**	**Volume**
Climate Var.	**Spring Mean Temp**	**Cold Season Precip**	**Climate Var.**	**Spring Mean Temp**	**Cold Season Precip**
# of Stations (%)	**18 (51%)**	**16 (46%)**	**# of Stations (%)**	**18 (51%)**	**16 (46%)**
10BE001	-0.39		07OB001		
10CD001			07NB001	No data	No data
10ED002	-0.42		Fort Providence	-0.59	
07FD002		0.36	10GC001	-0.72	
07GJ001		0.54	10KA001	-0.67	
07KC001			10LC014	-0.62	
07AD002	-0.49	0.49	10MC002	-0.63	
07DA001		0.33	08BB005	-0.49	0.62
07JD002			08CE001	-0.69	0.51
07LE002			08CG001	-0.64	0.45
05DF001			08DB001	-0.61	0.60
05GG001		0.33	08EF001	-0.59	0.64
05AJ001			08FF001		
05CK004			08FA002	-0.47	0.53
05HG001	No data	No data	08FB006	-0.54	0.74
05KJ001	(0.33)		08FB007	-0.53	0.63
08MF005			08FC003		0.68
08NM085		(-0.30)	08GD004	-0.47	
08NE058		(-0.33)	**R_{avg}***	**-0.43**	**0.41**

*Excluding values in brackets which are opposite in magnitude to remaining R values for each variable.

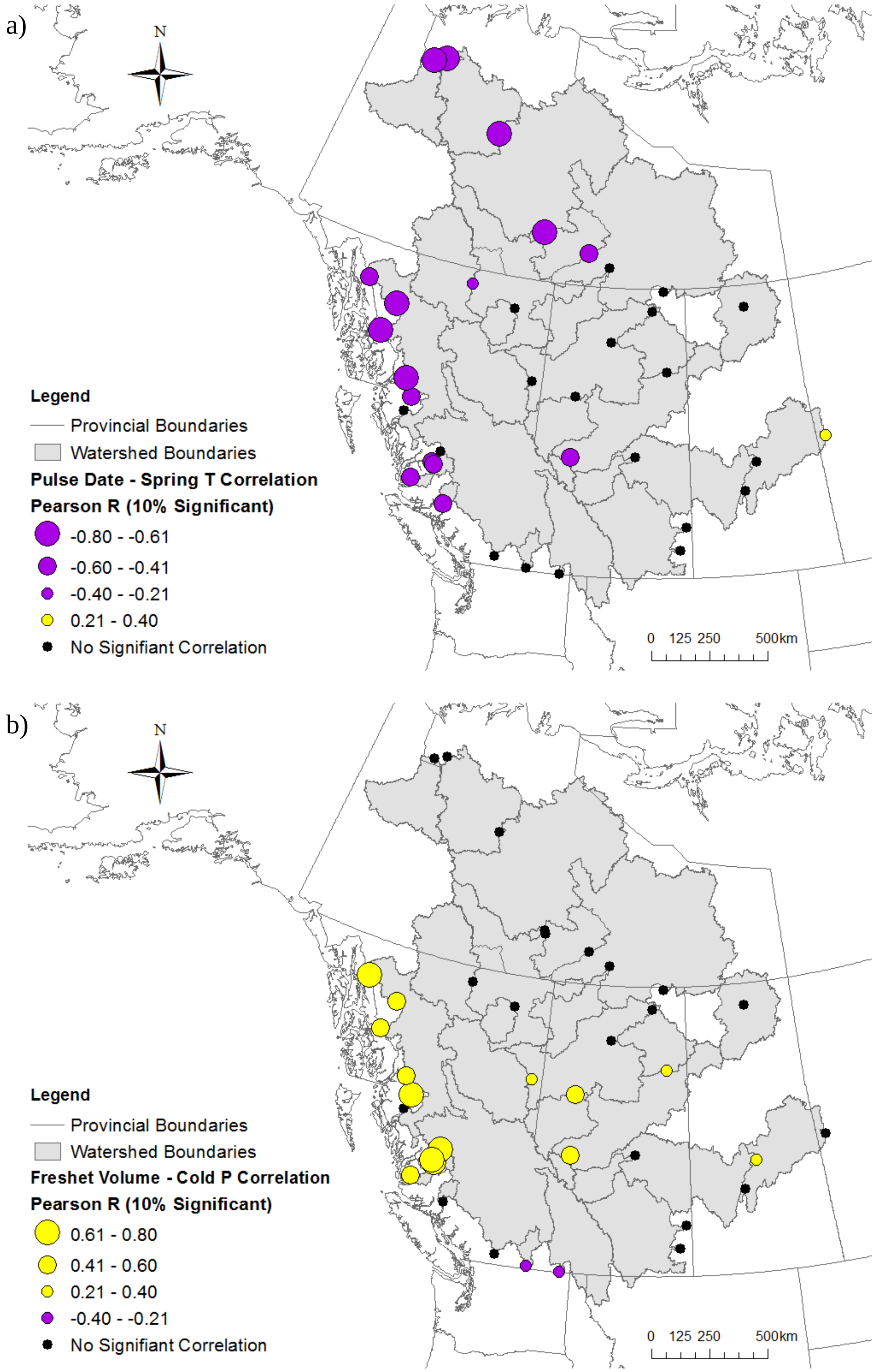


Figure 5.21. Stations that exhibited significant relationships between a) pulse date and spring mean temperature, and b) freshet volume and cold season precipitation.

It is clear that the Pacific coast region was most sensitive to changes in temperature and precipitation between 1976-2010. Earlier spring freshets in rivers located along the west coast, as well as in rivers situated in the northern parts of the study region, were shown to be highly correlated with spring warming (as verified by results of pulse date and spring temperature trend analysis). In addition, increasing freshet volumes in the North Pacific basin were shown to be highly correlated with increases in precipitation in the same region.

5.3.2 Teleconnections with Large-Scale Atmospheric Processes

As discussed in *Section 2.2.4.2*, large-scale atmospheric and oceanic processes can exert different effects on land surface variables affecting hydrology when in different phases of the climate signal. Each climate index has a unique period of persistency for each phase, as outlined in *Section 4.1.3*. The periodicity of each climate index used in this study can be gauged through inspection of Figure 5.22, which plots climate index values for each signal from 1976-2010.

Climate signals generally persist over at least a season, rather than over a single month or less (e.g., Maurer *et al.*, 2004). For example, the ENSO pattern generally tends to exert the largest effect on climate during the winter months (Bonsal *et al.*, 2006; Bonsal, 2013, personal communication). For this reason, composite analyses between climate indices and runoff/freshet variables were performed using minimum 3-month climate index means. For consistency, seasonal runoff variables were analyzed with respect to seasonal climate index values up to four seasonal lags prior. The influences of winter, spring, cold season, and Bonsal Spring climate index means on freshet variables were also assessed.

Significant teleconnections between each runoff/freshet variable and each climate index are presented in Table 5.17. A relationship was deemed significant if at least six basins or nine stations experienced runoff/freshet timing (or volume) that differed significantly ($p < 0.1$) from the series mean for the eight years of highest (positive phase) or lowest (negative phase) climate index values. The number of basins/stations that exhibited a significant relationship between runoff/freshet and the climate index is given, as is the percent out of 25 watersheds/34 stations. The positive phase of a climate index is denoted by a “(+)”, while the negative phase is denoted by a “(–)”.The relationship between the hydrologic variable and the climate index phase is also given; a “lower” relationship implies that the mean runoff or freshet timing/volume for the eight years of highest or lowest climate index values was significantly lower than the series mean (as detected by the t-test). Example plots for two basins that exhibited a reduced runoff relationship with one phase of a climate index are given in Figure 5.23.

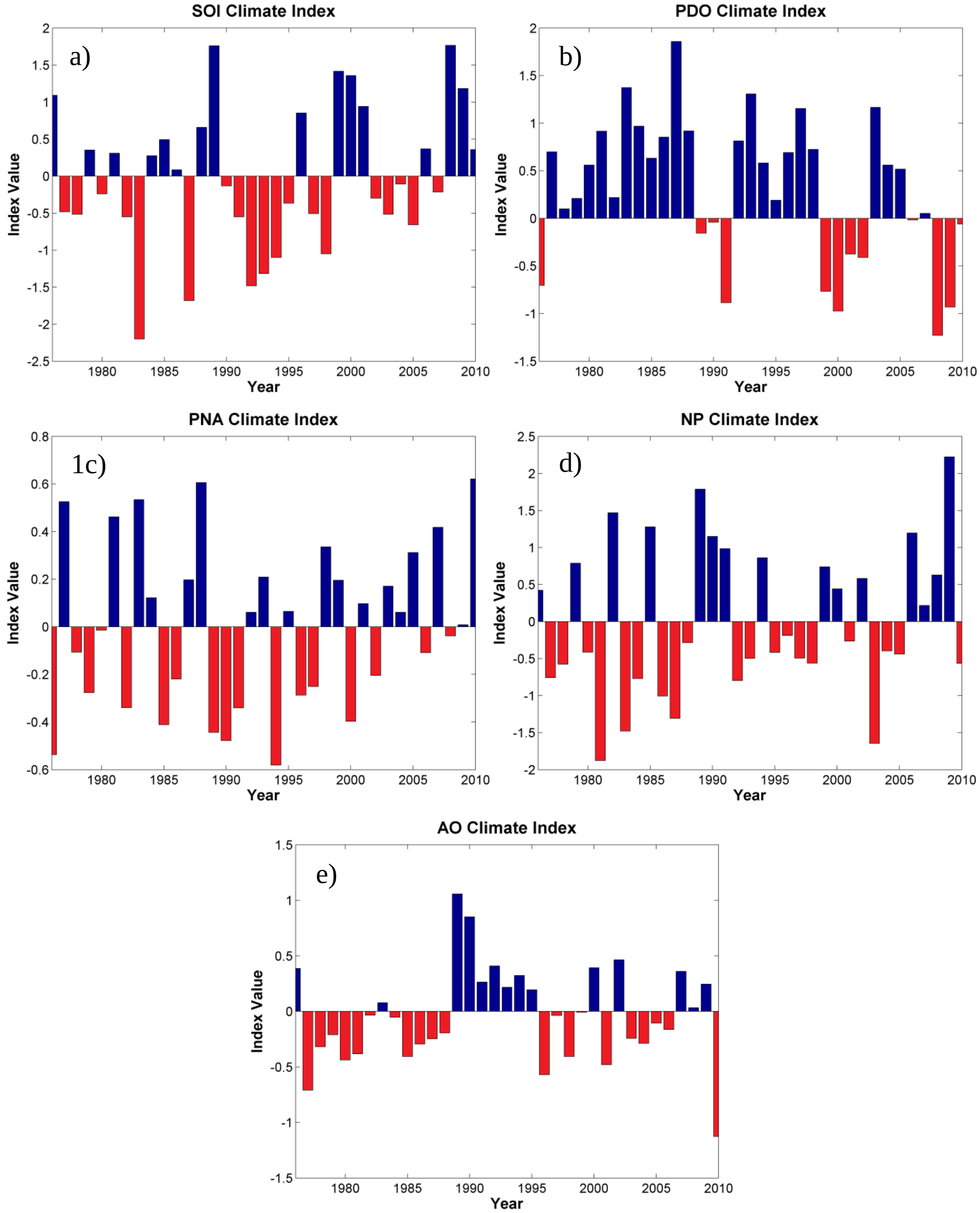
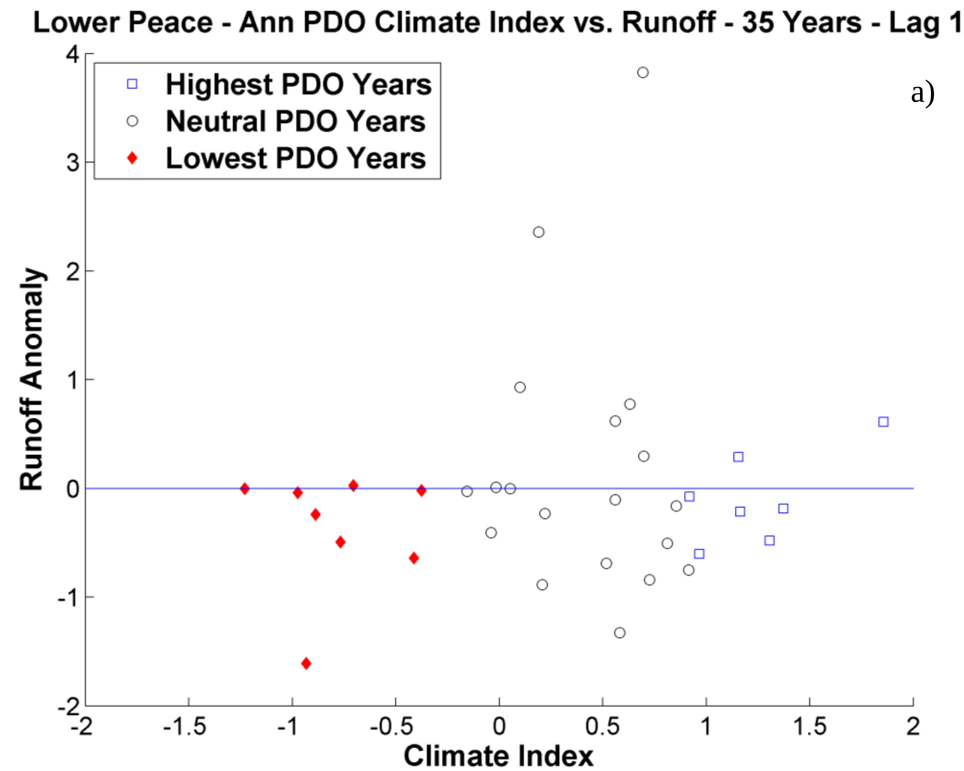


Figure 5.22. Time series plots of the a) SOI, b) PDO, c) PNA, d) NP, and e) AO climate patterns from 1976-2010.

Table 5.17. Summary of composite analysis between runoff/freshet variables and climate indices.

Variable	Climate Index	(Phase)	Season	Lag	# Basins/ Stations	% Basins/ Stations	Relationship*
Runoff							
Annual Runoff	PDO	(-)	Ann	1	12	48%	Lower
Winter Runoff	PDO	(-)	Summer	2	6	24%	Lower
	AO	(+)	Spring	3	7	28%	Lower
	PNA	(-)	Spring	3	7	28%	Lower
	NP	(-)	Winter	4	6	24%	Lower
	PNA	(+)	Winter	4	9	36%	Lower
Spring Runoff	PDO	(+)	Winter	1	6	24%	Lower
	PNA	(+)	Winter	1	7	28%	Lower
	PNA	(+)	Fall	2	7	28%	Lower
	SOI	(-)	Fall	2	7	28%	Lower
	PNA	(+)	Summer	3	8	32%	Lower
Summer Runoff	AO	(-)	Winter	2	7	28%	Lower
Fall Runoff	AO	(-)	Summer	1	6	24%	Lower
	AO	(+)	Spring	2	9	36%	Lower
	PDO	(+)	Fall	4	6	24%	Lower
	PDO	(-)	Fall	4	9	36%	Lower
Cold Season Runoff	PNA	(+)	Warm Season	1	6	24%	Lower
	PDO	(-)	Warm Season	1	9	36%	Lower
	NP	(-)	Cold Season	2	6	24%	Lower
Warm Season Runoff	PDO	(+)	Cold Season	1	6	24%	Lower
Freshet							
Pulse Date	-	-	-	-	-	-	-
CM Date	-	-	-	-	-	-	-
Length	NP	(+)	Bonsal Spring	-	11	30%	Lower
	NP	(+)	Cold Season	-	9	24%	Lower
Volume	PNA	(+)	Bonsal Spring	-	9	24%	Lower
	PNA	(+)	Winter	-	9	24%	Lower
	PNA	(+)	Cold Season	-	11	30%	Lower

* Average runoff/freshet variable anomaly with respect to mean zero for the eight years associated with the significant teleconnection (climate index/phase)



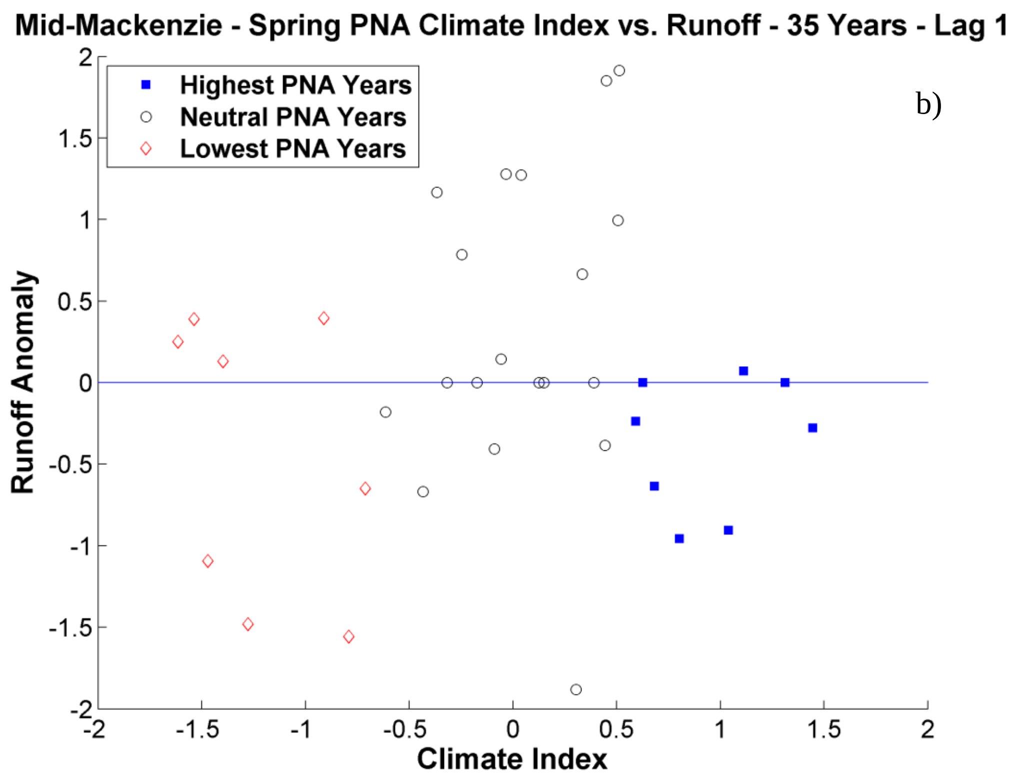


Figure 5.23. Scatterplots of a) annual PDO versus runoff anomaly of the following year, and b) winter PNA versus runoff anomaly of the following spring.

Results of composite analysis revealed that the negative phase of the PDO and the positive phase of the PNA patterns were significantly linked to river runoff in western Canada for the period of 1976-2010. Figure 5.24 shows a map of basins that exhibited a significant relationship between annual runoff and the negative phase of the PDO from the preceding year, as well as a map of the stations at which freshet volume was significantly influenced by the positive phase of the cold season PNA. While the SOI and AO patterns did exhibit certain links with runoff, these patterns were not as clear or consistent as the PDO and PNA patterns. The positive phase of the NP pattern during the cold season and Bonsal Spring was shown to influence the length of the spring freshet by shortening this significant hydrologic event, however neither it nor any other climate pattern was shown to cause changes in the timing of the onset or ending of the freshet.

The negative phase of the PDO from the preceding year resulted in reduced annual runoff in 9 out of the 12 basins that exhibited a significant relationship. Only the Peel, Upper Mackenzie, and Lower Liard River basins experienced increased runoff due to the negative PDO; all three of these basins are located north of 60°N. Winter and cold season runoff were significantly linked to the negative phase of the PDO of the preceding summer and warm season, respectively. Again, all basins that displayed a significant relationship experienced reduced winter/cold season runoff during the eight lowest years of the summer/warm season PDO index, with the exception of the Lower Mackenzie River basin for winter runoff, also located within the NR. Furthermore, fall runoff was significantly reduced the year following a fall negative PDO event, except in the Lower Mackenzie and Peel River watersheds. Overall, runoff during the cold months were shown to be linked to negative PDO events from the preceding warm months (6 to 12 months prior), where the negative PDO resulted in reduced runoff in the mid- and lower-latitudes, but increased runoff in the North.

The link between runoff and the negative phase of the PDO was surprising considering the PDO was in a positive phase for the majority of the study period (i.e., from 1977-1998; see Figure 5.22b). Moreover, given that it was the negative phase that exhibited a significant teleconnection with runoff, it is interesting that reduced runoff still resulted in the majority of basins during the years of most negative PDO (since the negative phase of the PDO is generally associated with cooler, wetter conditions, and hence increased streamflows). The direct link between cooler, wetter conditions and increased runoff was only apparent over the northern-most basins, while the opposite effect was shown in the mid- and low-latitudes. This implies that the trends and patterns noted in western Canadian river flow between 1976-2010 were not primarily controlled by the influence of the PDO (i.e., natural variability), except perhaps in the North. This result stands in contrast to the strong, negative relationship between streamflow and the PDO discovered by St. Jacques *et al.* (2010), but likely speaks to the short study period used here (versus at least 90 years).

The positive phase of the PNA was shown to significantly influence runoff following the onset of the spring freshet. Both spring runoff and total freshet volume were significantly reduced following a positive winter PNA event; spring runoff was also influenced by a positive PNA during the preceding fall and summer, while freshet volume was connected with both Bonsal Spring and cold season positive PNA events. These results reflect the warm, El Niño-like conditions that are associated with a positive PNA event – less snowpack is likely to accumulate during a warmer winter, while less precipitation and more evaporation could be expected during a warmer summer and fall, both of which would result in lower-than-normal freshet and spring flows. Cold season runoff was also shown to be significantly influenced by a positive PNA event during the preceding warm season, whereas runoff one year following a positive winter PNA event was also shown to be reduced.

Interestingly, however, the majority of stations that exhibited a significant relationship between decreased freshet volume and the positive phase of the PNA actually experienced increasing (or lack of) trends in freshet volume (as well as spring and cold season) overall. This again shows that natural variability (teleconnections) was not a dominant controller of the trends shown in runoff volume (or streamflow timing).

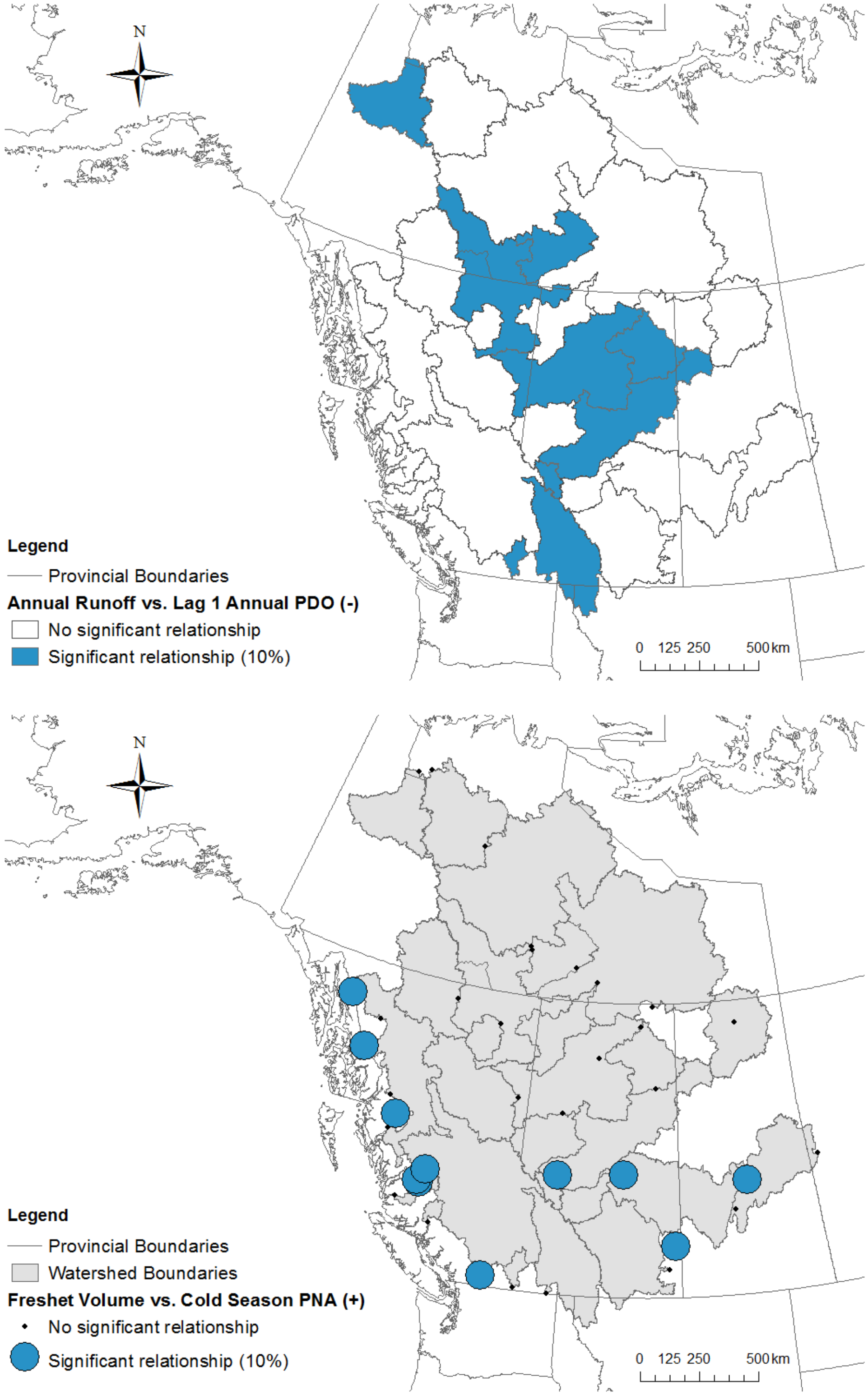


Figure 5.24. a) Basins that exhibited a significant relationship between annual runoff and the negative phase of the PDO from the preceding year. b) Stations that exhibited a significant relationship between freshet volume and the positive phase of the cold season PNA.

Chapter 6

Conclusions and Future Work

The redistribution of western Canadian water resources and the link to changing climate was examined through an analysis of trends and patterns in streamflow volume and timing in 25 conterminous western Canadian watersheds. This chapter details the overall conclusions drawn from the results of the study. *Section 6.1* provides a discussion of the results, drawing on references from the literature and implications for water resources managers in western Canada. Recommendations for areas of further research on this topic are offered in *Section 6.2*.

6.1 Discussion of Results

Section 2.1.2.1 outlined many of the recent trends and patterns detected in western Canadian hydrology. In particular, literature results stressed the tendency for increased runoff in the North (Bates *et al.*, 2008), especially during the cold seasons; decreased runoff in the semi-arid Canadian Prairies and in warm season runoff derived from the Rocky Mountains (e.g., Barnett *et al.*, 2005; Schindler and Donahue, 2006; St. Jacques *et al.*, 2010); and contrasting decreases and increases in summer and winter flows, respectively, along the Pacific coast of BC (Whitfield and

Cannon, 2000). In addition, earlier spring freshets have been detected in watersheds across western Canada (Stewart *et al.*, 2004), with the exception of the Pacific basins (Whitfield and Cannon, 2000). These results have been driven by increased evaporation, especially in the North, as well as a gradient of precipitation change, including enhancement in the North and decreases in the South (Serreze *et al.*, 2000).

The results of this study have shown general agreement with the conclusions of previous research, including a general trend toward increased runoff in the North and related decreased flow in the mid-latitudes since at least 1976 (and since 1966 for basins south of 60°N). The most southern basins of the study region, however, did not show consistent drying as has been otherwise reported. Specifically, the South Saskatchewan River basin experienced slight increases in runoff between 1976-2010, despite reports of long-term decreases in streamflow in this watershed (e.g., Schindler and Donahue 2006); this result is likely due to the short period of study, as longer term (i.e., 45-year) results do indicate decreases, at least during the warm season. Hydrological coherence among each of these regions – namely, the northern region (NR), mid-latitude region (MR), and southern region (SR) – was strongest annually and during the warm month seasons. Climatic drivers of observed trends were not consistent among each region.

Amplified climate change in the high-latitudes has caused continuous decreases in snow and ice cover, northward migration of precipitation patterns, more profound sea ice melting, glacial wastage, permafrost degradation, and increases in runoff in Arctic watersheds (Polyakov *et al.*, 2002; Serreze and Francis, 2006). For the CROCWR study region, increases in high-latitude runoff were most consistent during the winter and fall, when all basins north of 60°N exhibited increasing (or a lack of) trends in runoff. Increases in traditionally very low winter flows in the Arctic have also been related to greater subsurface infiltration into thawing permafrost stocks (ACIA, 2005).

During the warm seasons and annually, trends in the NR were not as uniform as they were during the cold months. Specifically, the Peel and Mid-Mackenzie watersheds did not exhibit increasing trends as did the other basins north of 60°N. Important to note, however, is that despite prior studies indicating increases in northern precipitation, this study has shown that only the North

Pacific basin and the Fort Nelson basin, a small sub-watershed of the Liard River basin, exhibited (non-significant) increases in 1976-2010 annual precipitation, while the remainder of basins experienced decreases in annual precipitation. Furthermore, the Peel and Mid-Mackenzie River basins actually experienced significant ($p < 0.05$) decreasing trends in precipitation, which likely drove the decreases in runoff noted in these basins (despite a lack of significant Pearson correlation between runoff and precipitation in the Peel watershed). Given the overall opposite trend directions, precipitation was not statistically correlated with runoff over most of the NR. Reported increases in precipitation are concentrated further north than the reaches of the CROCWR study region (Linton *et al.*, 2013).

During spring, warming was strongest over the NR, when the Peel River and three Mackenzie River basins all exhibited significant, high magnitude increasing trends in mean temperature. Warming of the NR caused more rapid snowmelt and consequentially earlier spring freshets over this region. Northern river freshet volumes also increased, in line with increases in warm season runoff. Trends in northern runoff in the CROCWR region appeared to be more affected by trends in temperature than by changing precipitation, although neither measure of climate showed a significant correlation with runoff in all or most of the NR basins. Increased northern river flows were also influenced by the cool, wet negative phase of the PDO.

Increases in northern river streamflows are especially important when considering the mid-latitude origins of these rivers. Significant drying was observed over the majority of the MR, implying a northward shift in water availability from adjacent southerly basins. As discussed, however, increases in high latitude CROCWR runoff were shown to have not been driven by increases in precipitation, therefore other factors (i.e., temperature/evaporation related) are more likely to have caused the wetter NR.

In the mid-latitudes, decreases in runoff were shown to be strongly connected to both increases in temperature and decreases in precipitation; hence, the cause-and-effect relationship between climate and water in this region was strong. The most significant decreasing trends in runoff tended to concentrate in and around the Lower Athabasca River basin. Interestingly, the onset of the spring freshet was shown to have generally become later in this region, except in rivers along the Pacific coast. This is likely related to decreasing spring temperatures detected among many of the eastern mid- and low-latitude basins, although Pearson correlations between spring air temperature and onset of the spring freshet were not significant in this region. Spring freshet volumes were shown to have significantly decreased, similar to overall and warm season runoff results.

The North and South Pacific basins displayed increasing trends in runoff annually and during most seasons. Runoff in the North Pacific basin, in particular, was shown to be highly influenced by changing precipitation, similar to other basins in the MR, but in the opposite direction (i.e., the North Pacific basin experienced increased precipitation and runoff, while the majority of the MR exhibited decreases in both). Not surprisingly, spring freshet volumes were also shown to have increased along the coast, and were strongly linked with cold season precipitation.

Except for the South Saskatchewan River basin, the Rocky Mountains-sourced SR experienced mainly decreasing trends in runoff during all seasons, with the exception of spring, when all basins exhibited increasing (or null) trends in runoff for the 35-year period. While neither temperature nor precipitation trend directions were consistent among all SR basins for spring, spring runoff in all SR basins did correlate significantly with precipitation from the previous fall. Spring runoff was also shown to be related to the positive phase of the Pacific North American (PNA) climate pattern, but only during years when runoff was below average (which does not explain why increased spring runoff resulted overall). Interestingly, the tendency of increased runoff in the SR was not present over the 45-year period; rather, all southern basins exhibited decreased (or lack of change in) spring runoff over the period of 1966-2010.

Overall, trends appear to have been driven by a combination of near-surface climate, atmospheric patterns (to a lesser extent), and other external factor(s) (including human impact and regulation). Specifically, runoff in the MR was strongly connected to changes in both temperature and precipitation. The negative phase of the PDO was also shown to have some effect on decreased runoff in a number of basins in the MR.

Increased runoff in the NR was not shown to be associated with either temperature or precipitation, though changes in climate have no doubt been causal in these results. Unfortunately, low quality hydrometric and climate data for northern Canadian watersheds could have potentially complicated these results. In addition, the lack of consistency in trend direction, specifically with reference to the Peel and Mid-Mackenzie basins, reinforces this point, but also represents another complicating factor in northern hydrology: that storage of water is difficult to account for. Specifically, release of water from storage (i.e. melt of snow, permafrost) could not be quantified through this analysis. This fact could have some influence on the results shown.

Runoff in the SR was shown to be significantly linked to precipitation, though the directions (i.e., increasing/decreasing) of runoff and precipitation trends did not always align. Again, there were complicating factors in the South, mainly due to the degree of regulation that has taken place within these basins (e.g., the South Saskatchewan and Fraser River basins). Human impact from heavy water use, diversion, and storage, especially in southern Alberta and Saskatchewan, has obscured the natural hydrology of some of these rivers, hence fluctuations in regulated gauge records reflect both global warming effects and direct human impacts (St. Jacques *et al.*, 2010). Unfortunately, while hydrologic modelling makes it possible to quantify losses to evaporation and to determine the naturalized flow response of regulated rivers, this was outside of scope of this particular work. Given the relationship between runoff and precipitation, coupled with the projections of climate models which forecast further reduced precipitation in the lower latitudes, it is likely that water availability in the SR will decrease in the future, even without considering the expected increases in water demands of a growing economy and population (St. Jacques *et al.*, 2010).

In conclusion, both spatial and temporal changes were detected in western Canadian river flow. These results have important implications to water resources managers in western Canada, especially considering projected increases in future high latitude runoff and contrasting drying of the lower-latitude Prairies. Future runoff-regime patterns and will likely create a greater need for revised water management, including a potential for re-allocation of water resources.

6.2 Future Research

Although western Canada has been targeted as an area of hydrologic concern due its unique vulnerability to climate change on account of its large alpine water storage, dry continental interior, and large south-to-north and west-to-east river systems, it is nevertheless important to examine the impacts of climate change on water resources in other regions of Canada. The east coast, for example, has been shown to have cooled over at least the last 50 years; whether this will transcribe into increased or decreased water security should be explored. In addition, as implied through this work, precipitation changes in the North are occurring at even higher latitudes than the CROCWR study region. It would be interesting and potentially alarming to investigate flows in the higher latitudes, given that increased runoff has already been observed over northwestern Canada even in the absence of increased precipitation (in fact, in the presence of drying!). Similar studies over regions including central Canada and southwestern Ontario may be of assistance in assessing the role of climate change versus human impacts (especially in densely populated and highly developed areas, such as southwestern Ontario) on Canadian water resources. Finally, investigation of river flow trends to the west of the CROCWR study region – i.e., in Alaskan rivers – may help to further understand/quantify the role of the Aleutian low on northwestern streamflow.

The calculation of streamflow/runoff data for each CROCWR basin using WSC (output – input) stations is an inexact measure of watershed productivity. Although this method should account for the majority of runoff generated within each watershed, it does not distinguish inputs or outputs of precipitation, evaporation, snowmelt, other tributary river flows, or other releases from/gains to storage. Furthermore, watershed storage components and volumes are not defined, an important factor especially in basins where large lakes are located. This is particularly complicating in the northern basins, where large lakes and permafrost store large quantities of water, as well as in the South, where regulation of rivers makes determining the natural effect of climate change on water resources difficult. More hydrologically meaningful water balances can be solved using hydrologic modelling software, such the Variable Infiltration Capacity (VIC) model (http://www.hydro.washington.edu/Lettenmaier/Models/VIC/), for example.

With respect to the timing of the onset of the spring freshet, it would be useful to examine the relationship between pulse dates and the retreat of the spring 0°C isotherm (Bonsal and Prowse, 2003). Although likely to be affected in all regions that experience "winter", the most prominent changes in the 0°C isotherm are expected to occur at the high latitudes, where warming has been shown/projected to be strongest.

In terms of relating near-surface climatic variables to hydrology, relationships could be more fully explored to emphasize the role of multi-variable effects (e.g., the combined effect of temperature and precipitation). Such analyses could be accomplished through parametric methods, such as Multiple Linear Regression (Bawden *et al.*, 2013) or Generalized Least Squares regression (St. Jacques *et al.*, 2010), in which the sensitivity or "elasticity" of streamflow volume or timing to each measure of climate can be assessed. Similarly, it would be useful to develop a relationship between streamflow and the major atmospheric circulation pattrns controlling streamflow variability in western Canada, as one could then factor out the effects of "natural variability" on streamflow data. In the absence of these confounding factors, streamflow trends would reflect true hydroclimatic changes (i.e., from global warming) as well as the influence of human impact (St. Jacques *et al.*, 2010).

Apart from delineating statistical linkages between atmospheric circulation modes and streamflow, it is critical to note that such linkages offer little insight unless they are readily explained by physical phenomenon or help generate new hypotheses in the case of apparently contradictory results (Li, 2012). The role of synoptic patterns and their relationship with atmospheric modes and resulting near surface climatic variables (i.e., temperature and precipitation) will help to further explain the influence of teleconnections on CROCWR runoff (Newton, 2013). The specific role of the Aleutian low should also be investigated, due to its presence in most western North American teleconnections. Beyond teleconnections, it would be useful to investigate the role of atmospheric moisuture transport over western and northern Canada, and how moisture is advected onto the continent (Zhang *et al.*, 2012). This information could feed into a more detailed assessment of changes in western Canadian precipitation (Linton, 2013), which in turn would relate to changes in water resources.

Finally, it is clear that the period of study plays a critical role in a trend analysis. By examining data from different time periods, one can arrive at very different conclusions. Notable differences in trend may even exist when a study period differs by as little as ten years (as this study has shown), hence a very different conclusion could be drawn if dealing with data sets of, say, 100 years or more. Though shorter term records can still lead to important decisions for shorter term water resources planning, it is important to consider the long range when planning for the long range future. Especially when assessing needs and planning infrastructure, longer hydrological records are important to evaluate the responses of river discharge to a range of natural climatic conditions – knowledge that is essential for informed management of water resources (Wolfe *et al.*, 2008). Unfortunately, there is a serious lack of both hydrological and climatic data in northern Canada. While longer-term analyses are possible using hydrologic/hydraulic modelled data, greater uncertainties will always exist (versus using actual collected data). Paleohydrologic and paleoclimate reconstruction data can provide long-term estimates of past hydrology and climate, respectively, but may downplay the effects of more recent non-stationarity in hydoclimatology (Milly *et al.*, 2008). Given the results of this research, as well as the projections from numerous other hydroclimatic studies with implications to northern Canadian hydrology, water monitoring agencies in northern Canada should consider making the enhancement of hydrometric station networks a priority – especially given the need to decipher the role of climate change/variability and for water managers to effectively respond to it.

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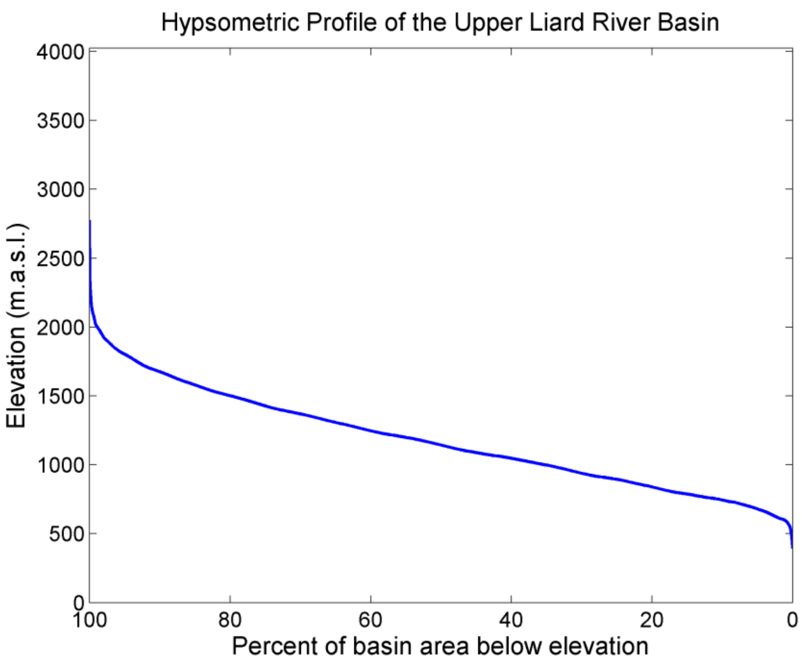
Appendices

Appendix A

Supporting Study Region Data


Hypsometric Profile of the Upper Liard River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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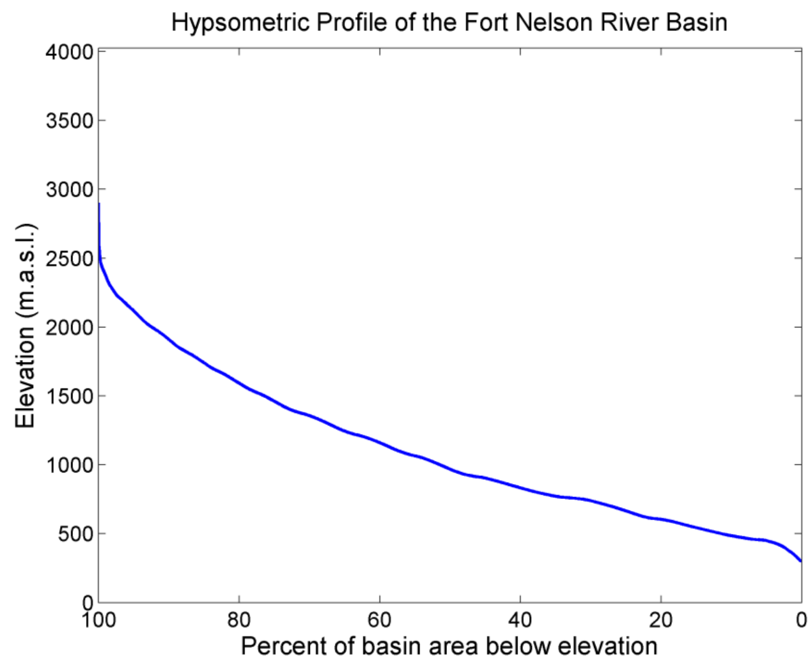
Hypsometric Profile of the Fort Nelson River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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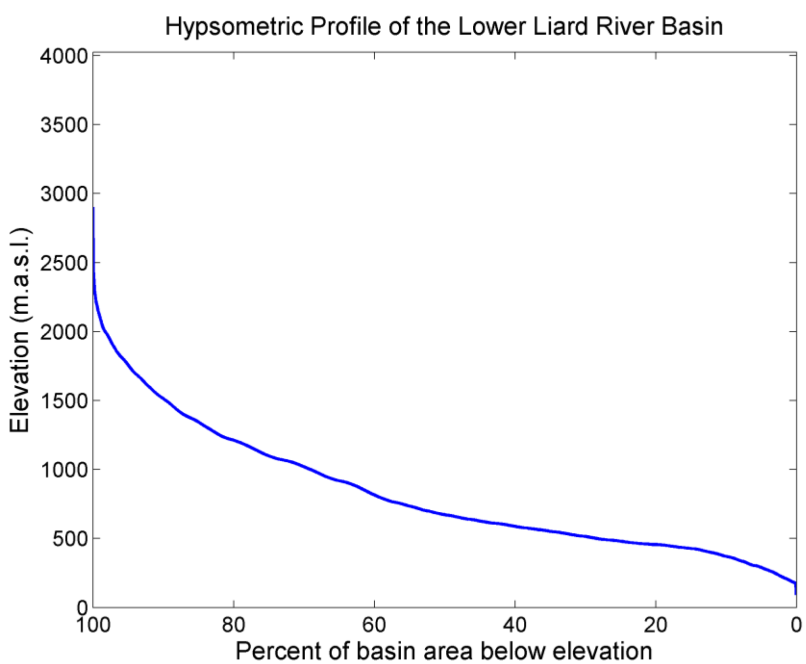
Hypsometric Profile of the Lower Liard River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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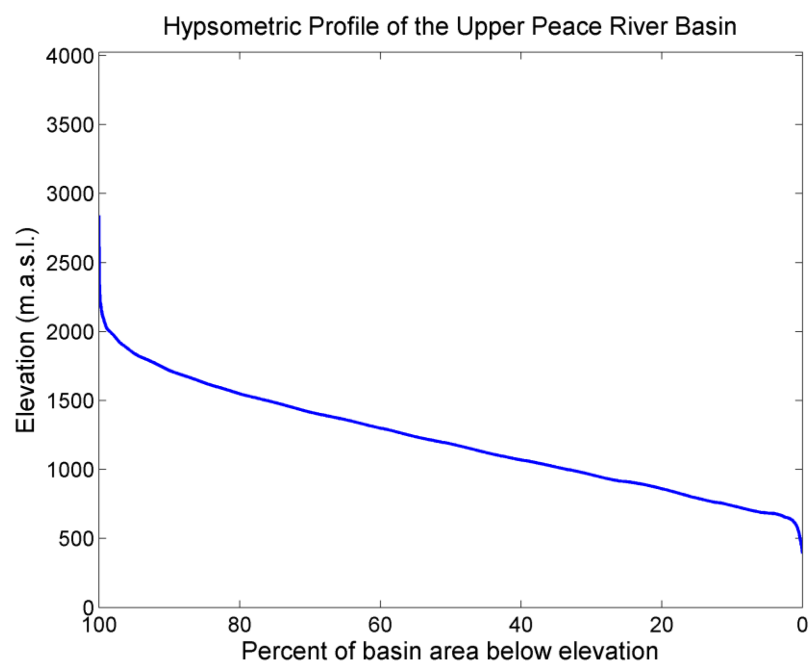
Hypsometric Profile of the Upper Peace River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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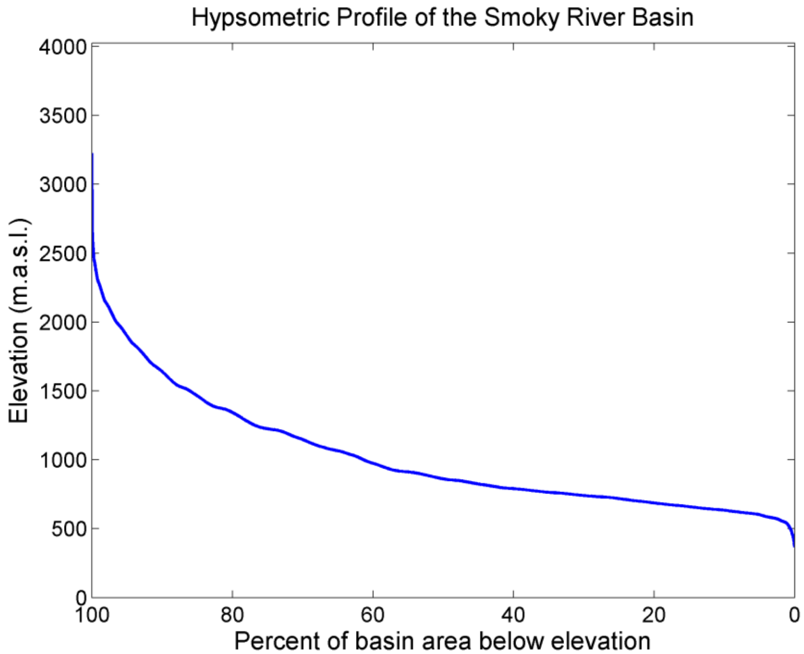
Hypsometric Profile of the Smoky River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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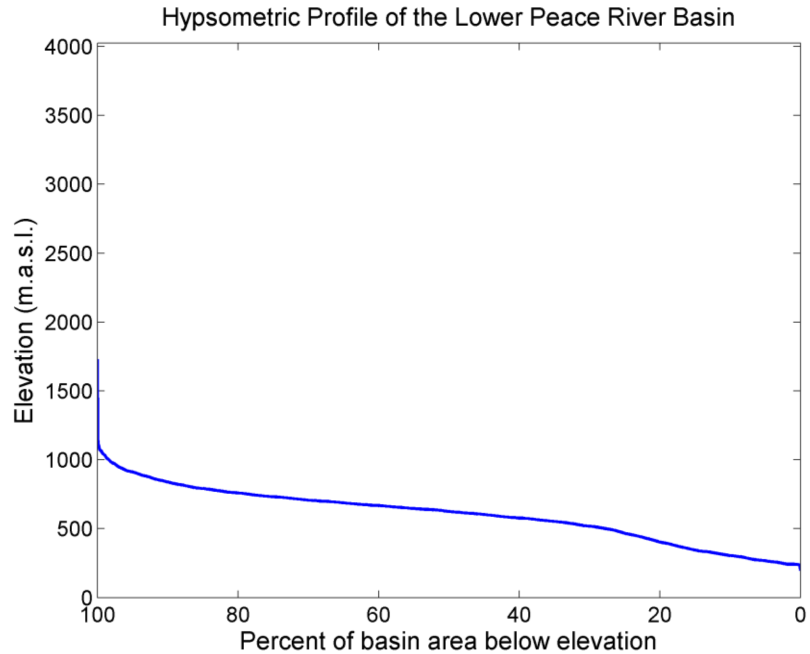
Hypsometric Profile of the Lower Peace River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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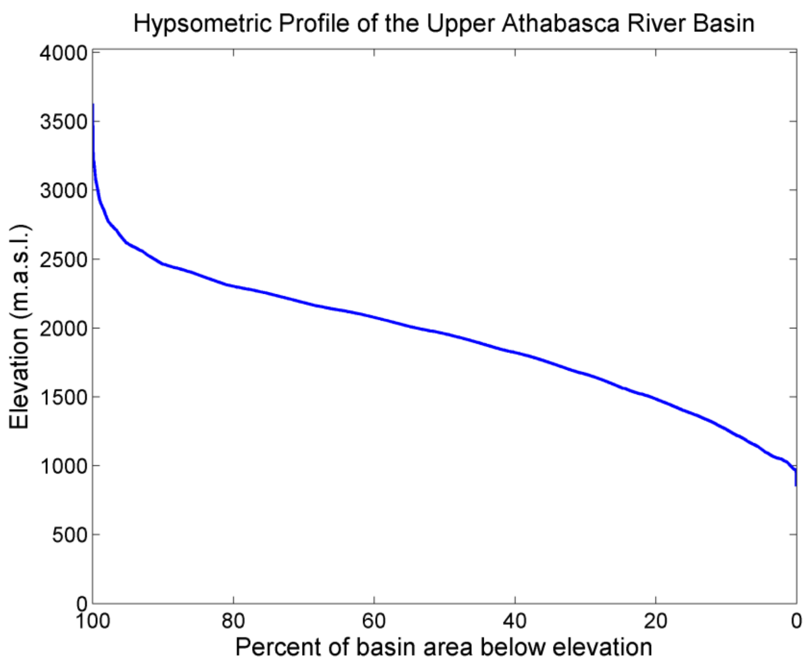
Hypsometric Profile of the Upper Athabasca River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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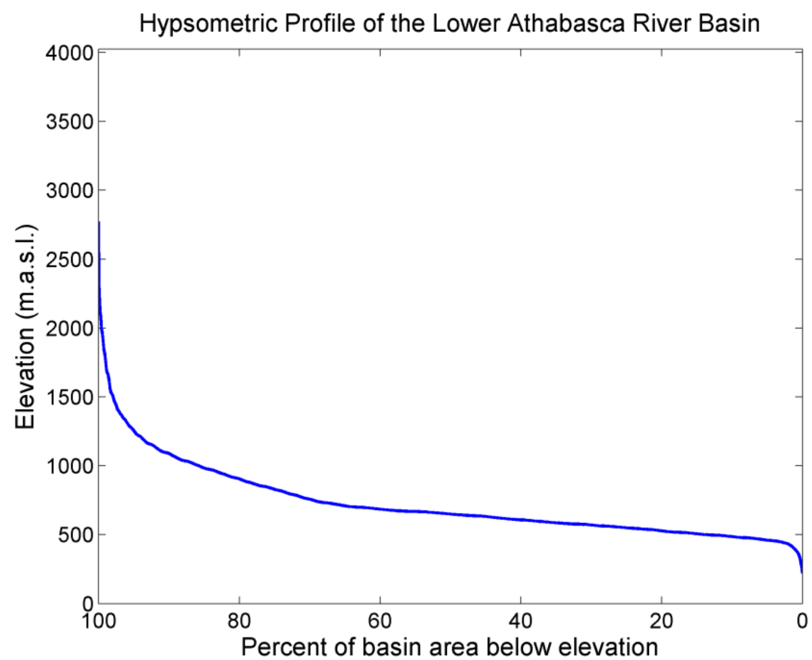
Hypsometric Profile of the Lower Athabasca River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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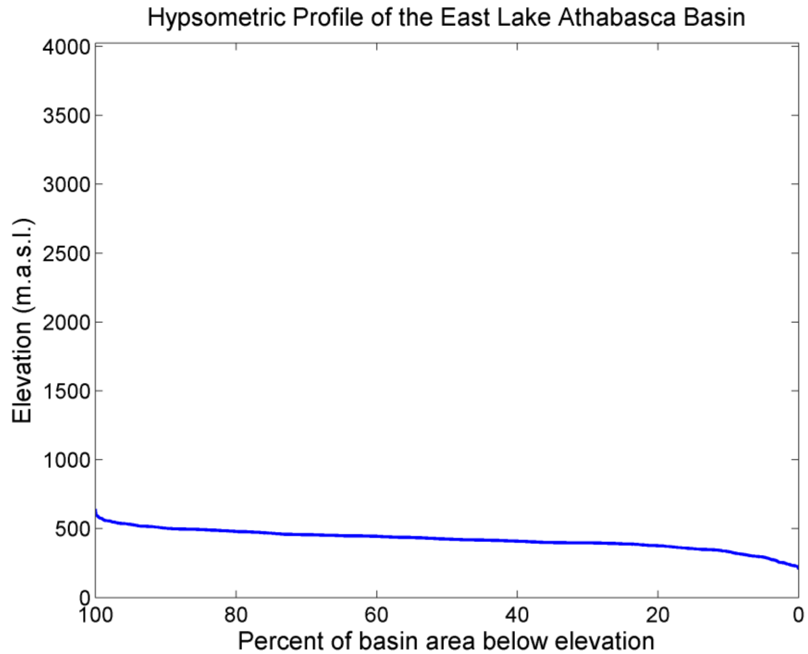
Hypsometric Profile of the East Lake Athabasca Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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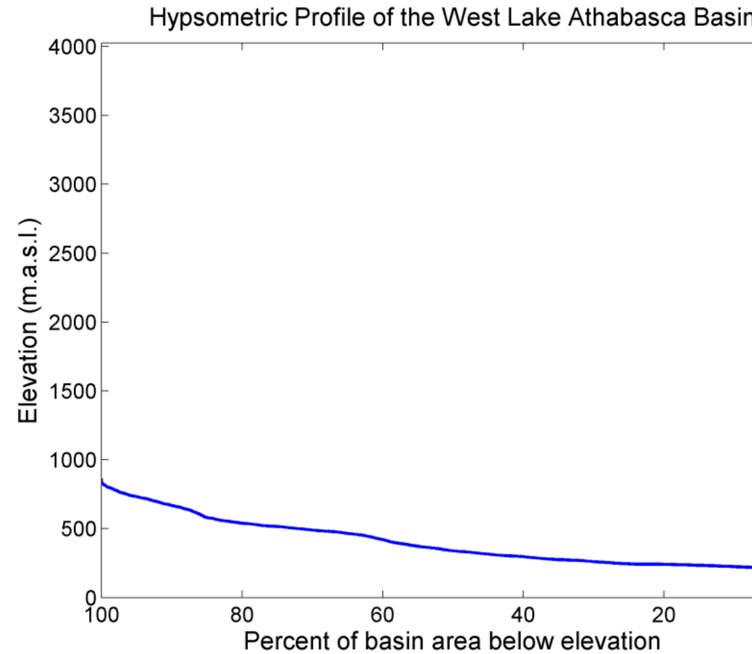
Hypsometric Profile of the West Lake Athabasca Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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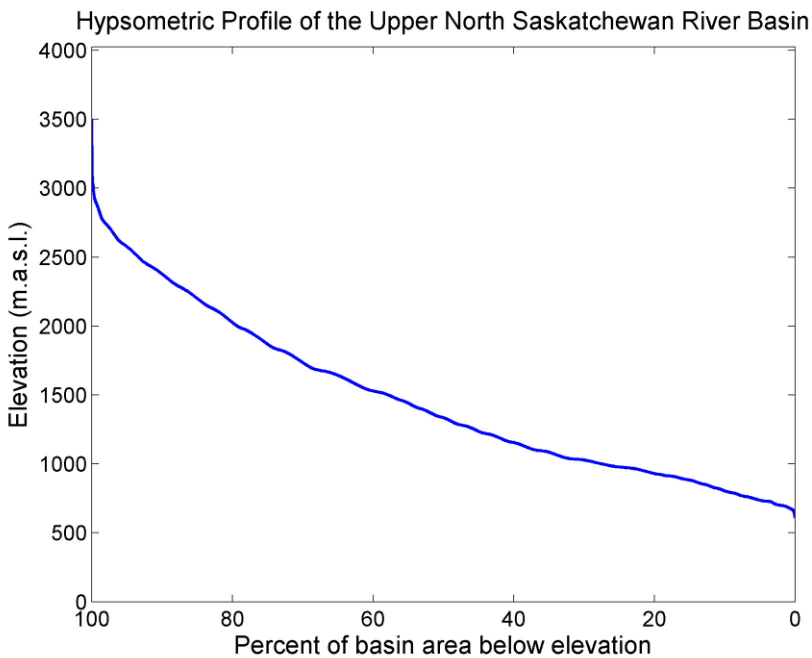
Hypsometric Profile of the Upper North Saskatchewan River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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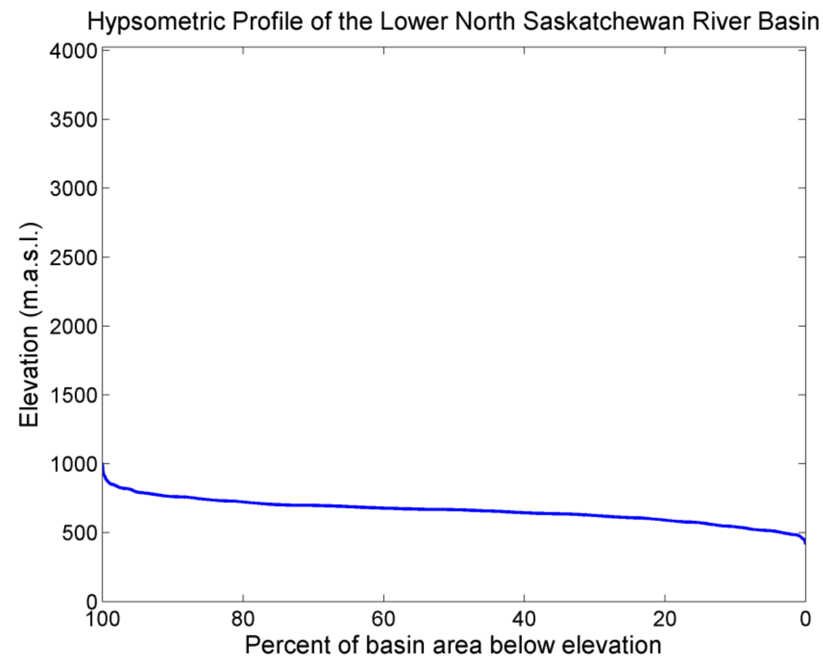
Hypsometric Profile of the Lower North Saskatchewan River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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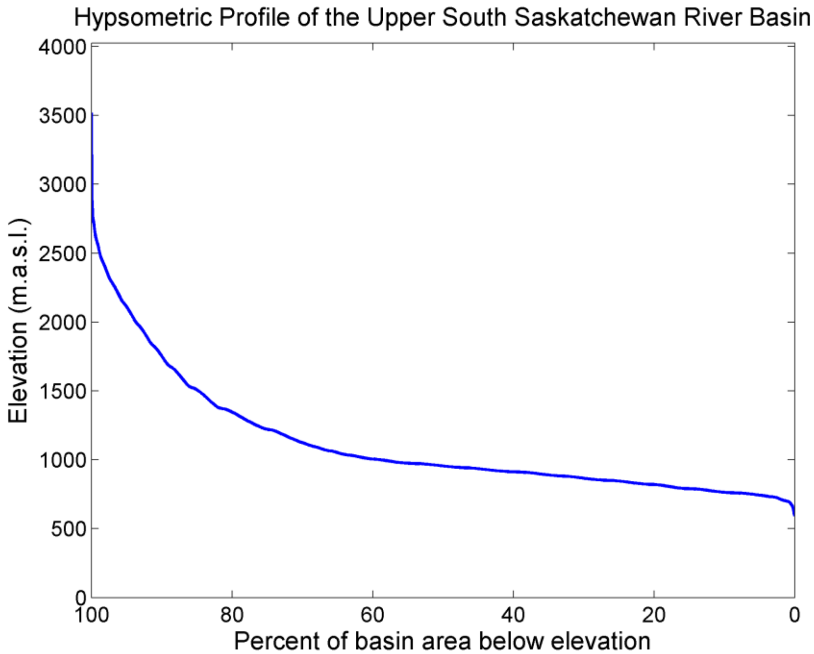
Hypsometric Profile of the Upper South Saskatchewan River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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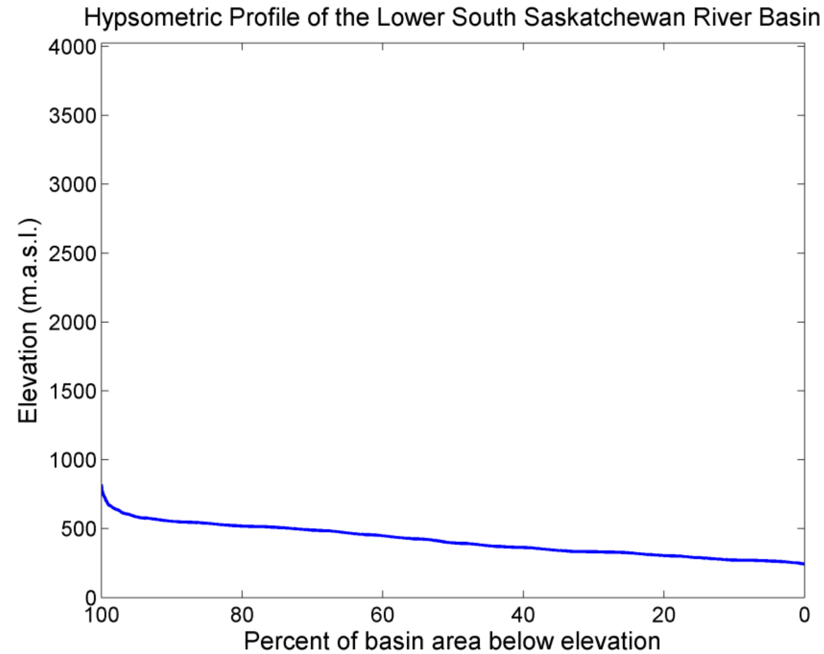
Hypsometric Profile of the Lower South Saskatchewan River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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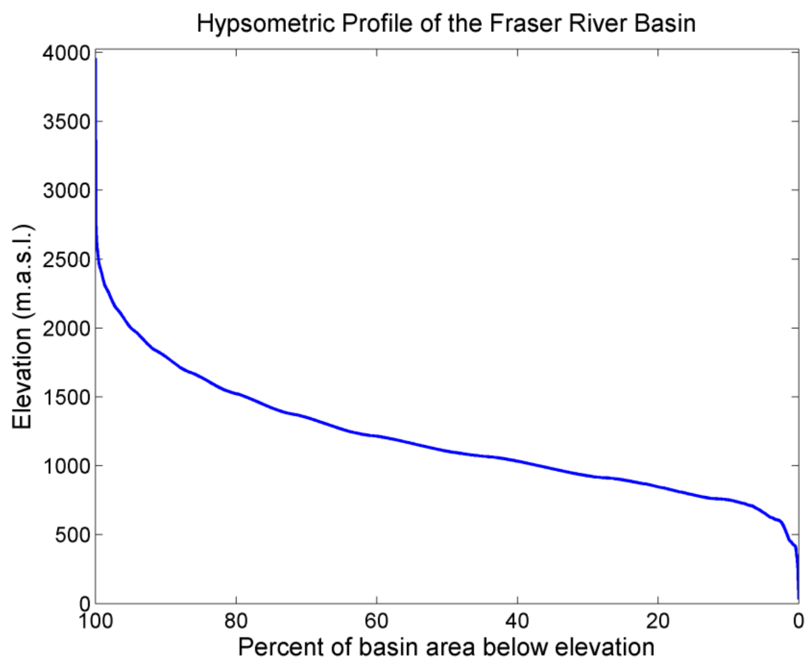
Hypsometric Profile of the Fraser River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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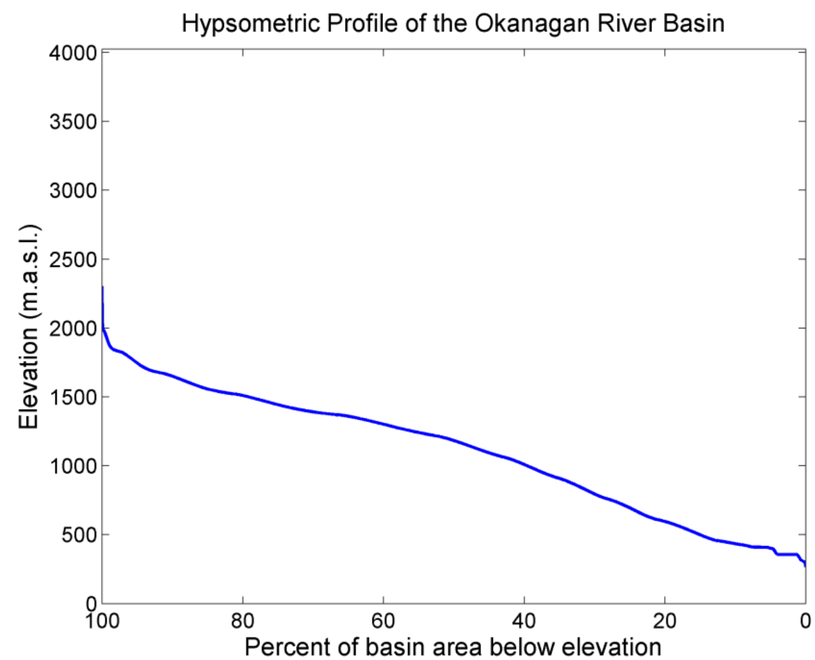
Hypsometric Profile of the Okanagan River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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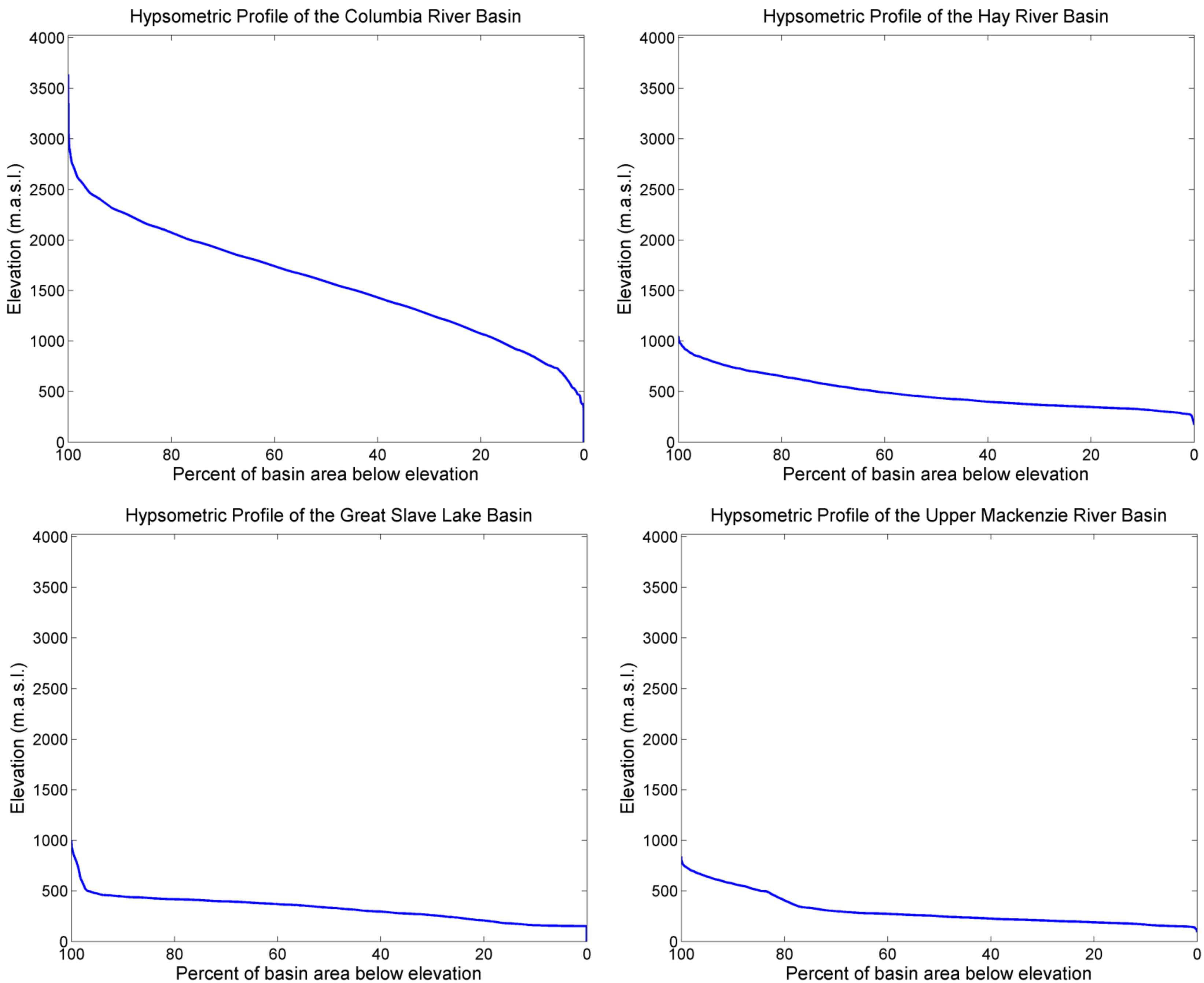
Hypsometric Profile of the Columbia River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
Hypsometric Profile of the Hay River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
Hypsometric Profile of the Great Slave Lake Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
Hypsometric Profile of the Upper Mackenzie River Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
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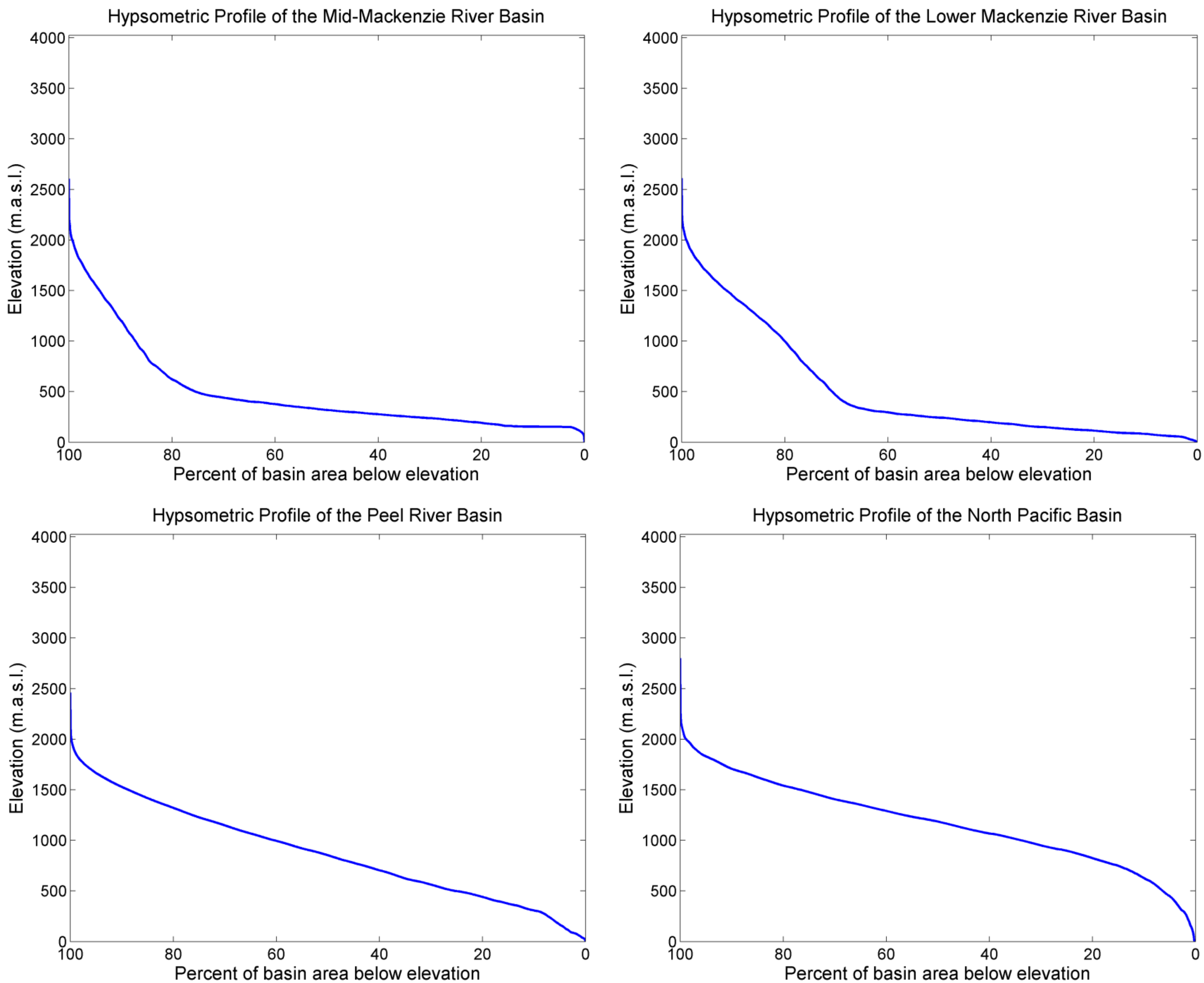

Hypsometric Profile of the Mid-Mackenzie River Basin
Hypsometric Profile of the Lower Mackenzie River Basin
Hypsometric Profile of the Peel River Basin
Hypsometric Profile of the North Pacific Basin
Elevation (m.a.s.l.)
Percent of basin area below elevation
0
500
1000
1500
2000
2500
3000
3500
4000
100
80
60
40
20
0

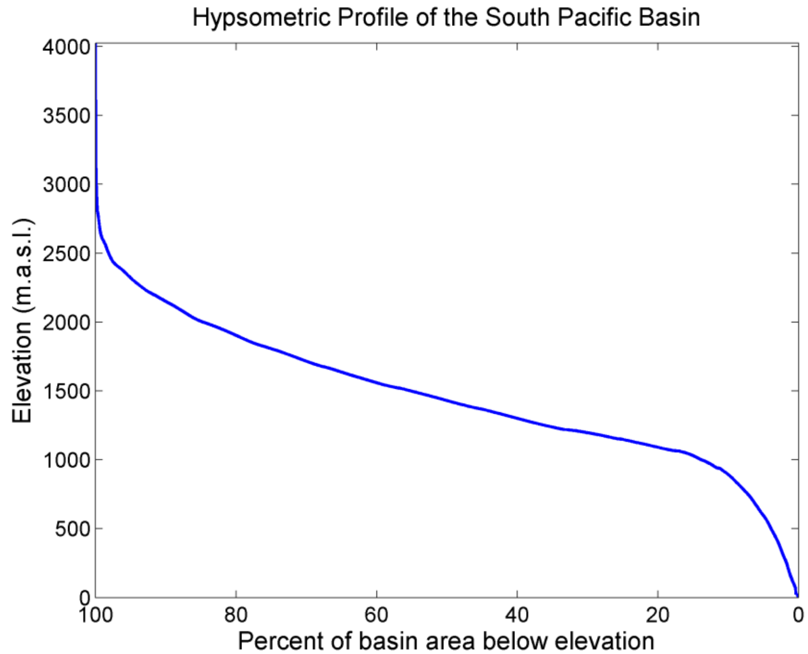


Figure A.1. Hypsometric profiles of the 25 CROCWR watersheds.

Table A.1. Mean annual streamflow and mean annual runoff values for the CROCWR study region for the period of 1966-2010.

Watershed		Annual Streamflow (m^3/s)	Annual Runoff (mm/yr)
1	Upper Liard	1135.1	345.2
2	Fort Nelson	212.8	332.3
3	Lower Liard	1153.2	242.4
4	Upper Peace	1416.0	442.3
5	Smoky River	325.3	205.0
6	Lower Peace	301.5	68.1
7	Upper Athabasca	169.7	551.0
8	Lower Athabasca	439.8	113.0
9	East Lake Athabasca	297.9	263.1
10	West Lake Athabasca	80.5	51.1
11	Hay	115.9	70.9
12	Great Slave	948.9	93.8
13	Upper Mackenzie	47.4	100.2
14	Mid Mackenzie	1655.1	164.3
15	Lower Mackenzie	664.7	240.9
16	Peel	683.1	307.7
17	North Pacific	3152.1	850.3
18	South Pacific	977.7	1583.9
19	Fraser	2711.8	395.8
20	Okanagan	18.0	75.0
21	Columbia	2801.3	570.0
22	Upper North Saskatchewan	195.9	220.4
23	Lower North Saskatchewan	33.6	10.4
24	Upper South Saskatchewan	218.3	66.2
25	Lower South Saskatchewan	138.3	37.3

Appendix B

Hydrologic Trends

Table B.1. Runoff trend results for the 1976-2010 (35-year) period.

Watershed	Annual (mm/35yrs)		Winter++ (mm/35yrs)		Spring (mm/35yrs)		Summer+ (mm/35yrs)		Fall++ (mm/35yrs)		Cold Season (mm/35yrs)		Warm Season++ (mm/35yr)	
	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend
Upper Liard	32.7	↑	6.7	↑↑	22.2	↑	-0.3	↓	9.1	↑	-53.6	↓↓	-9.9	↓
Fort Nelson	-64.6	↓	2.5	↑↑	-21.6	↓	-48	↓	-0.4	↓	-20.7	↓	-51.1	↓
Lower Liard	75.7	↑↑	2.2	↑	26	↑	22.8	↑↑	15.5	↑↑	5.3	↑	-87.7	↓
Upper Peace	-21.1	↓	21.3	↑	-17.9	↓	-23.5	↓	-16.6	↓	-5.1	↓	-29.8	↓
Smoky River	-58.9	↓↓	-1	↓	-27.5	↓	-34.7	↓↓	-1.6	↓	23.7	↑↑	11.6	↑
Lower Peace	-25.6	↓	-9	↓↓	1.8	↑	-3.2	↓	-10.6	↓	2.4	↑	-22.4	↓
Upper Athabasca	-25.5	↓	2.8	↑	1.9	↑	-22.9	↓	4.2	↑	-16.2	↓↓	-51.1	↓↓
Lower Athabasca	-60.3	↓↓	-4.7	↓↓	-15	↓	-28	↓↓	-11.4	↓↓	7.8	↑	49	↑
East Lake Athabasca	-43.4	↓	-3.9	↓	-13.3	↓	-21	↓	-8.2	↓	60	↑	150.5	↑
West Lake Athabasca	-29.2	↓↓	-1.3	↓	-6.9	↓	-9.2	↓	-2.7	↓	2	↑	-0.8	↓
Hay	7.6	↑	0.5	↑	-3.4	↓	1.6	↑	1.3	↑	-24.2	↓	-18.5	↓
Great Slave	59.3	↑↑	13.5	↑↑	24.6	↑↑	9.3	↑	8.1	↑	10.1	↑	8.4	↑
Upper Mackenzie	87.7	↑↑	0.7	↑	37.9	↑↑	43.6	↑↑	6.1	↑	0	None	-20.5	↓
Mid Mackenzie	-18.9	↓	4.3	↑	-5.5	↓	-12.7	↓	6.6	↑	2.2	↑	9.9	↑
Lower Mackenzie	300.3	↑↑	37	↑	57.3	↑↑	128.6	↑↑	89.5	↑↑	-11	↓	-17.5	↓
Peel	-10.2	↓	5.8	↑↑	-33.7	↓	15	↑	12.8	↑↑	11.7	↑↑	-26.7	↓
North Pacific	51.5	↑	8.7	↑↑	41.9	↑	12.7	↑	-11.7	↓	-5.3	↓	-74.2	↓↓
South Pacific	114.5	↑	8.6	↑	44.3	↑	9.9	↑	35.8	↑	11.9	↑	-10.4	↓
Fraser	-24.2	↓	-1.9	↓	2.2	↑	-19.2	↓	-0.3	↓	4.4	↑	-44.5	↓
Okanagan	-14.5	↓	-3.9	↓	1.2	↑	-4.6	↓	-4.3	↓↓	14.4	↑↑	9.4	↑
Columbia	-42.4	↓	-34.1	↓↓	3.9	↑	-0.2	↓	-9	↓	-8.8	↓	80.7	↑↑
Upper North Saskatchewan	-24.4	↓	-3.7	↓↓	-2.8	↓	-10.3	↓	-3.2	↓	-7.2	↓	-21	↓
Lower North Saskatchewan	1.4	↑	1.5	↑↑	0.5	↑	0.6	↑	0.3	↑	15.8	↑	-72.5	↓↓
Upper South Saskatchewan	13	↑	1.2	↑	7.7	↑	4.7	↑	3.6	↑↑	1.2	↑	10.1	↑
Lower South Saskatchewan	19.2	↑	0	None	3.8	↑	11.2	↑↑	7.1	↑↑	-3.8	↓	-26.8	↓↓

Table B.1(con't). Runoff trend results for the 1976-2010 (35-year) period.

Watershed	Annual		Winter[++]		Spring		Summer[+]		Fall[++]		Cold Season		Warm Season[++]	
	#	%	#	%	#	%	#	%	#	%	#	%	#	%
# ↓	11	44%	5	20%	10	40%	12	48%	10	40%	8	32%	13	52%
# **↓↓**	3	12%	4	16%	0	0%	2	8%	2	8%	2	8%	4	16%
# ↑	7	28%	9	36%	12	48%	7	28%	8	32%	11	44%	7	28%
# **↑↑**	4	16%	6	24%	3	12%	4	16%	5	20%	3	12%	1	4%
# No trend	0	0%	1	4%	0	0%	0	0%	0	0%	1	8%	0	4%
# Missing	0	0%	0	0%	0	0%	0	0%	0	0%	0	0%	0	0%

[+] Variable is field significant at 10% significance level.
[++] Variable is field significant at 5% significance level.
↑ or ↓: Non-significant trend
↑↑ or **↓↓**: Trend significant at 10% significance level
↑↑ or **↓↓**: Trend significant at 5% significance level

Table B.2. Runoff trend results for the 1966-2010 (45-year) period.

Watershed	Annual (mm/45yrs)		Winter^{++} (mm/45yrs)		Spring (mm/45yrs)		Summer^{+} (mm/45yrs)		Fall (mm/45yrs)		Cold Season^{+} (mm/45yrs)		Warm Season (mm/45yrs)	
	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend
Upper Liard	19.9	↑	9.7	↑↑	13.6	↑	-7.8	↓	9.6	↑↑	20.9	↑↑	9.1	↑
Fort Nelson	-13.2	↓	4.2	↑↑	-7.2	↓	-20.7	↓	4	↑	11.9	↑↑	-11.9	↓
Lower Liard	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Upper Peace	-3.9	↓	52	↑↑	-25.1	↓	-12.7	↓	25.7	↑	93.9	↑	-56.6	↓
Smoky River	-63.2	↓↓	-1.6	↓	-45.3	↓↓	-19.7	↓↓	-1.8	↓	-4.8	↓	-62.7	↓↓
Lower Peace	-5.8	↓	-6.1	↓	13.8	↑	-9.4	↓	-8.9	↓↓	-13.4	↓	-5	↓
Upper Athabasca	-60.1	↓	3	↑	-13.4	↓	-39.8	↓	3.6	↑	5.5	↑	-57.3	↓
Lower Athabasca	-66.6	↓↓	-3.6	↓↓	-20.1	↓↓	-23.7	↓↓	-9.3	↓↓	-7.4	↓	-49.2	↓↓
East Lake Athabasca	10.4	↑	3.3	↑	1.9	↑	-3.4	↓	-0.4	↓	3.8	↑	-15.2	↓
West Lake Athabasca	-28.3	↓↓	-0.3	↓	-10.9	↓↓	-7.8	↓	-2.3	↓	0.3	↑	-17.4	↓
Hay	25.5	↑	0.9	↑↑	1.5	↑	13.7	↑	5.9	↑↑	5.2	↑↑	11.7	↑
Great Slave	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Upper Mackenzie	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mid Mackenzie	-	-	-	-	-	-	-24.1	↓↓	-	-	-	-	-	-
Lower Mackenzie	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Peel	-	-	-	-	-	-	-	-	-	-	-	-	-	-
North Pacific	35.3	↑	16.6	↑↑	42	↑	-9.5	↓	0.1	↑	37.3	↑	52.3	↑
South Pacific	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Fraser	-53.6	↓↓	-3.1	↓	-8.8	↓	-28.5	↓↓	-7	↓	-2.1	↓	-48.7	↓↓
Okanagan	-4.6	↓	-2.1	↓	1.4	↑	2.9	↑	-1.5	↓	-6.4	↓	3	↑
Columbia	-81.2	↓	24.6	↑	-56.1	↓↓	-46.7	↓	13.8	↑	33	↑	-104.6	↓
Upper North Saskatchewan	-32.2	↓	5.9	↑	-16.2	↓	-29.2	↓	9.3	↑↑	16.9	↑↑	-47.7	↓↓
Lower North Saskatchewan	-0.3	↓	1.2	↑↑	0.4	↑	-1	↓	0.1	↑	1	↑	-1.1	↓
Upper South Saskatchewan	-18.4	↓	0.5	↑	-12	↓	-3.3	↓	2.3	↑↑	-3	↓	-10.9	↓
Lower South Saskatchewan	-	-	-	-	0.1	↑	2.6	↑	3.1	↑	-	-	-1.8	↓

Table B.2. Runoff trend results for the 1966-2010 (45-year) period.

Station	Annual		Winter++		Spring		Summer+		Fall		Cold Season+		Warm Season	
	#	%	#	%	#	%	#	%	#	%	#	%	#	%
# ↓	9	36%	5	20%	6	24%	12	48%	5	20%	6	24%	10	40%
# ↓↓	4	16%	1	4%	4	16%	4	16%	2	8%	0	0%	4	16%
# ↑	4	16%	5	20%	8	32%	3	12%	7	28%	7	28%	4	16%
# ↑↑	0	0%	6	24%	0	0%	0	0%	4	16%	4	16%	0	0%
# No trend	0	0%	0	0%	0	0%	0	0%	0	0%	0	0%	0	0%
# Missing	8	32%	8	32%	7	28%	6	24%	7	28%	8	32%	7	28%

+ Variable is field significant at 10% significance level.

++ Variable is field significant at 5% significance level.

↑ or ↓: Non-significant trend

↑↑ or **↓↓**: Trend significant at 10% significance level

↑↑ or ↓↓: Trend significant at 5% significance level

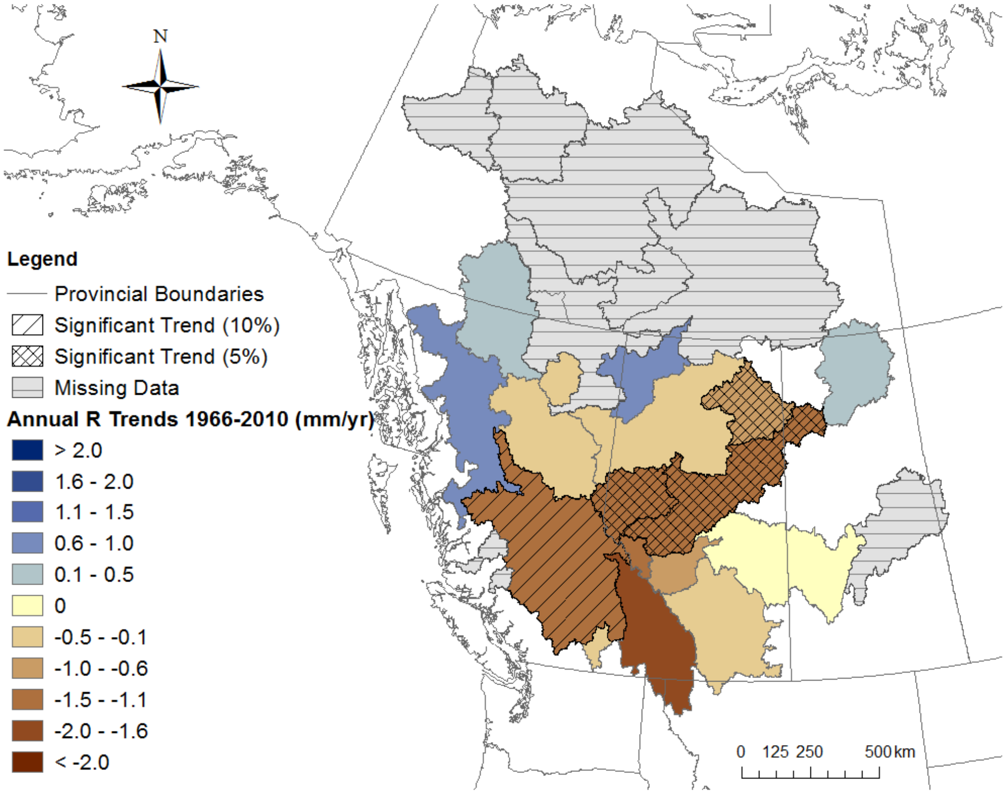
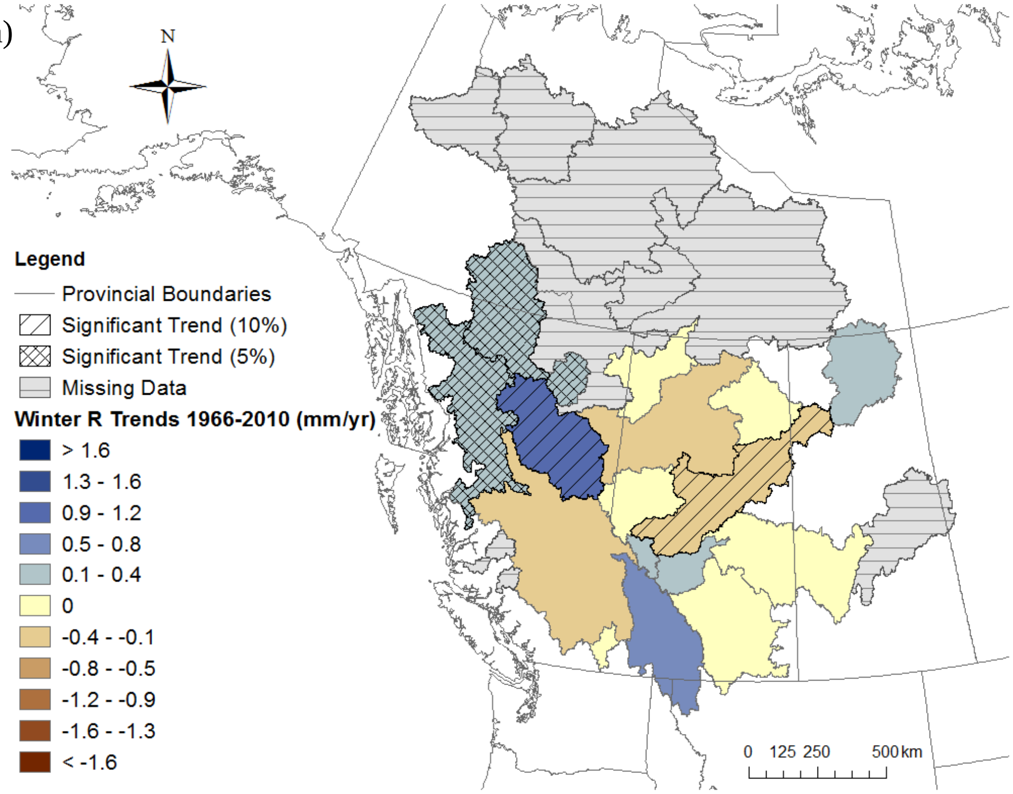


Figure B.1. Map of annual runoff trend slopes for the 1966-2010 analysis period. Basins exhibiting significant trends are shown as hatched.

a)



b)

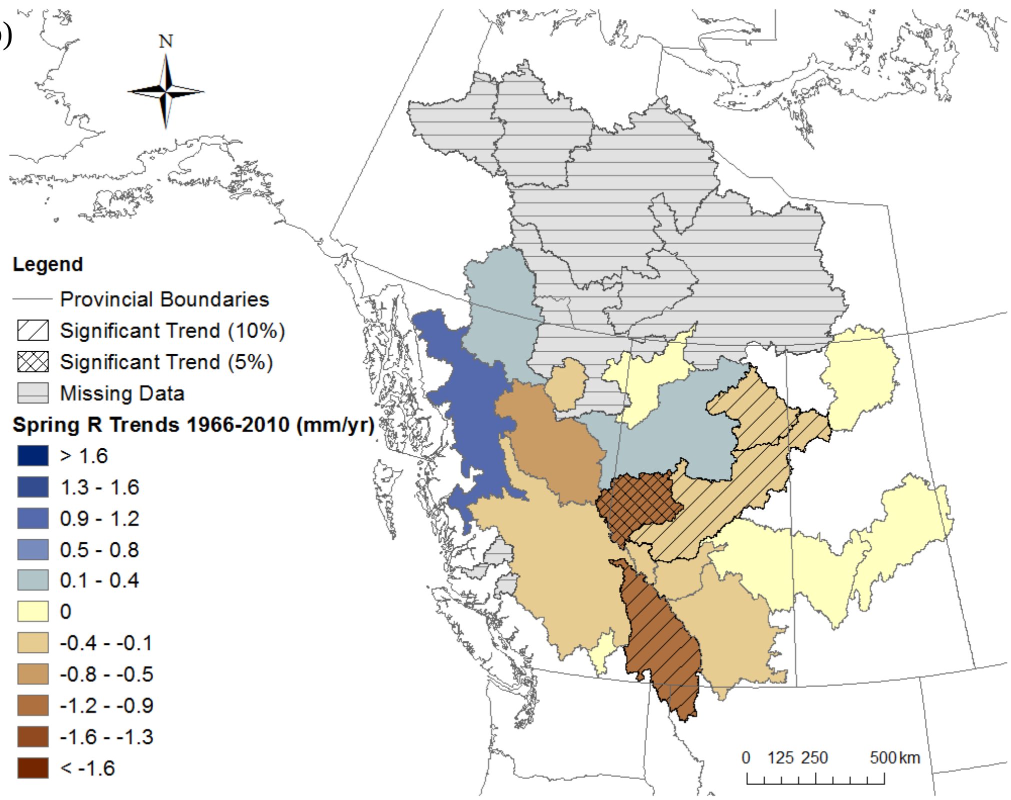
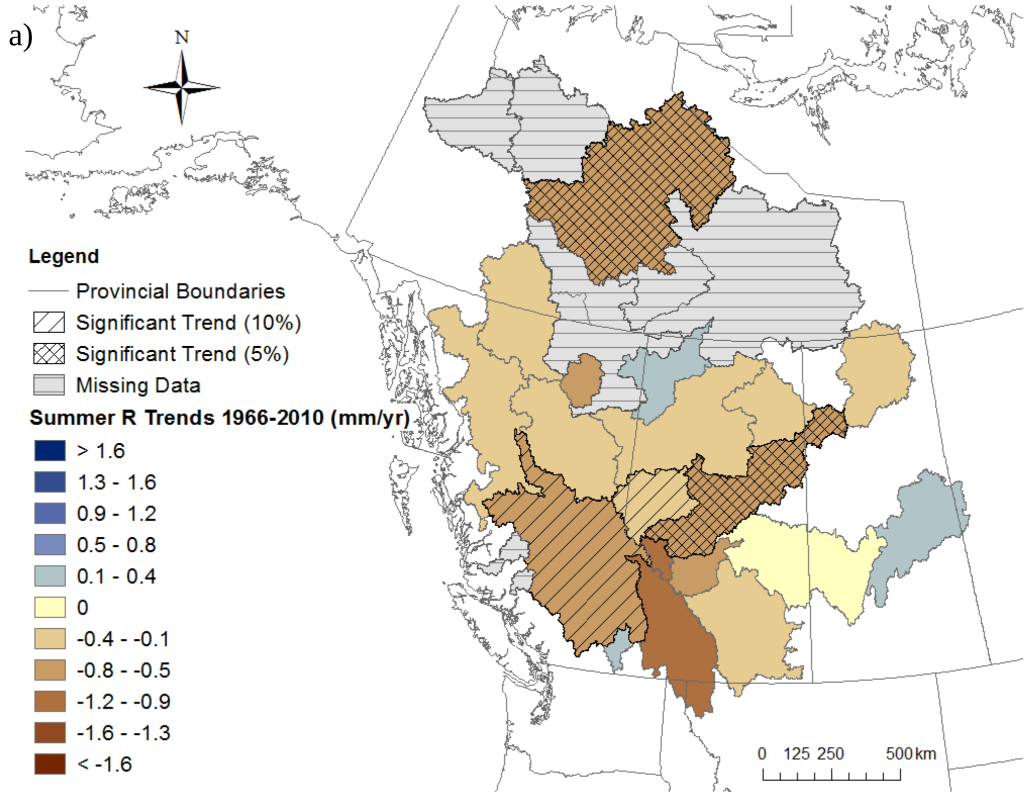


Figure B.2. Same as Figure B.1 but for a) winter and b) spring. Note that scale differs from Figure B.1.



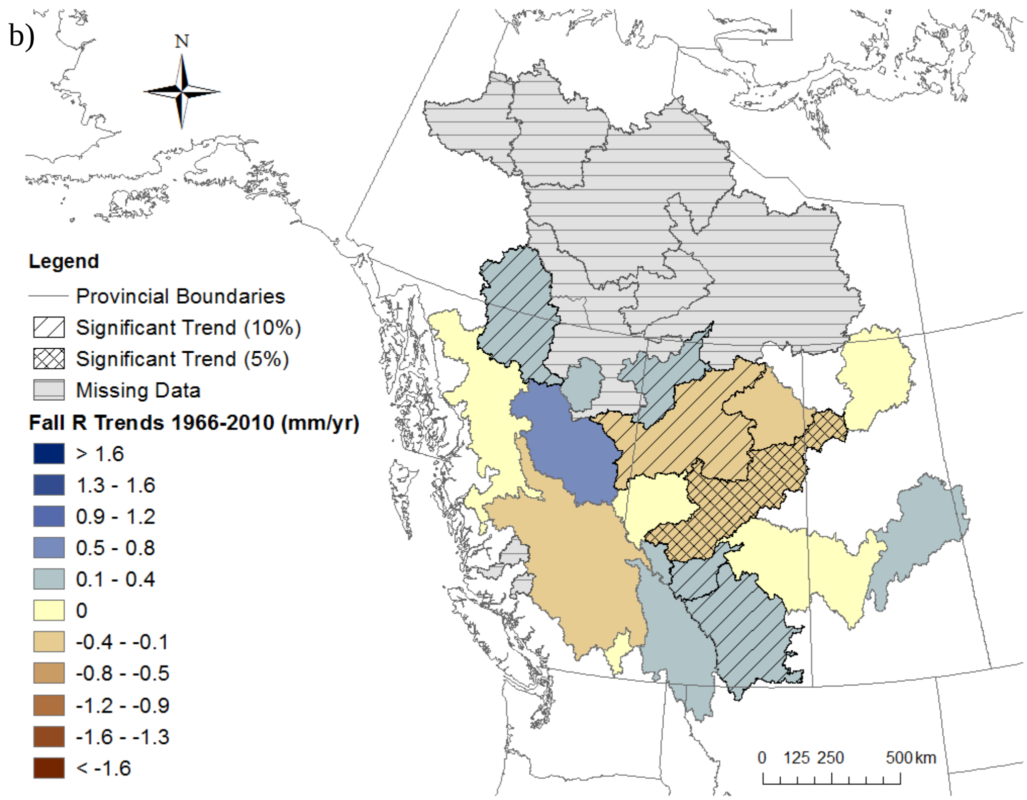
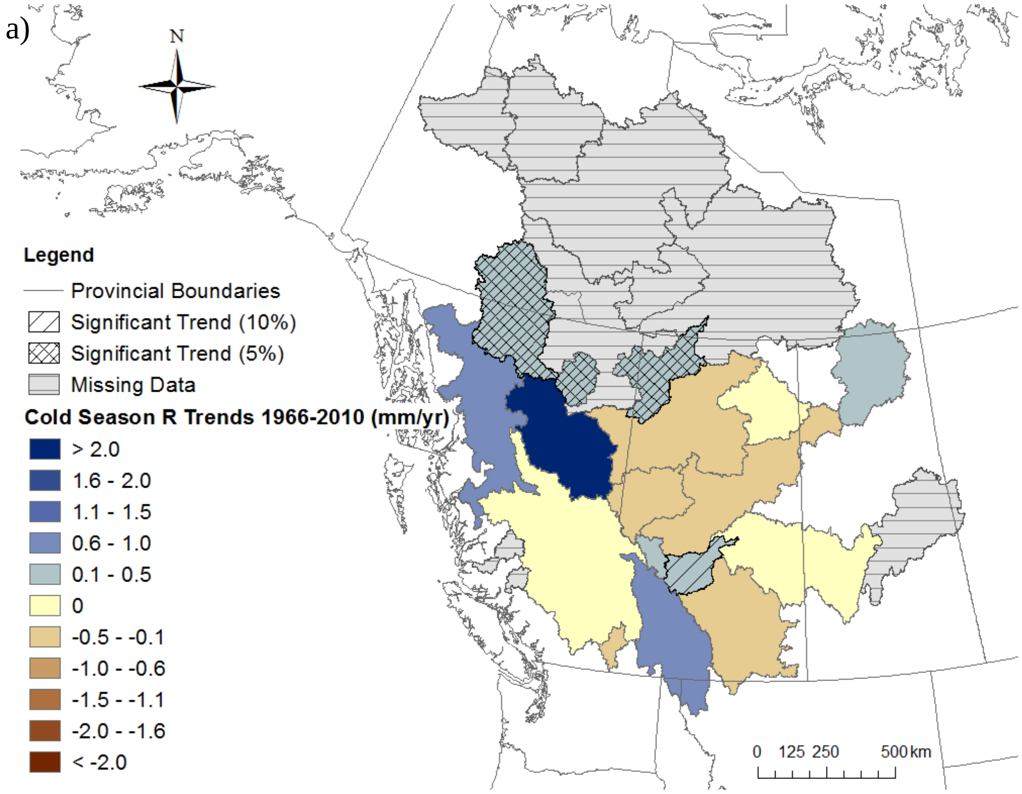


Figure B.3. Same as Figure B.1 but for a) summer and b) fall. Note that scale differs from Figure B.1.



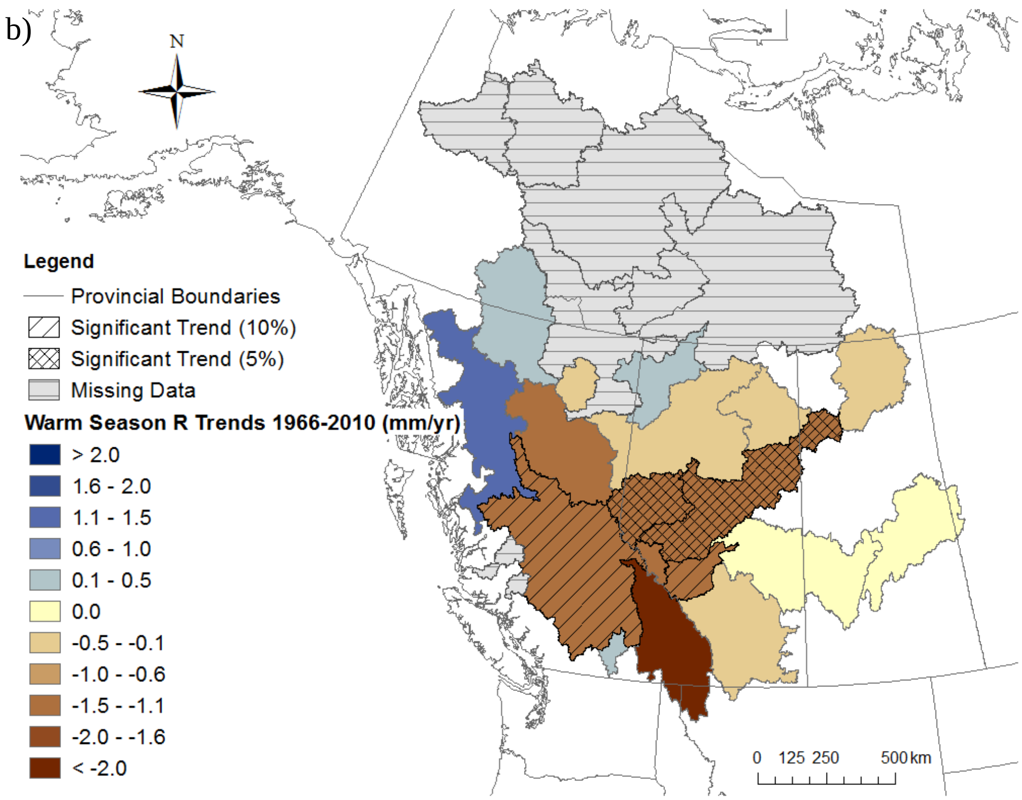
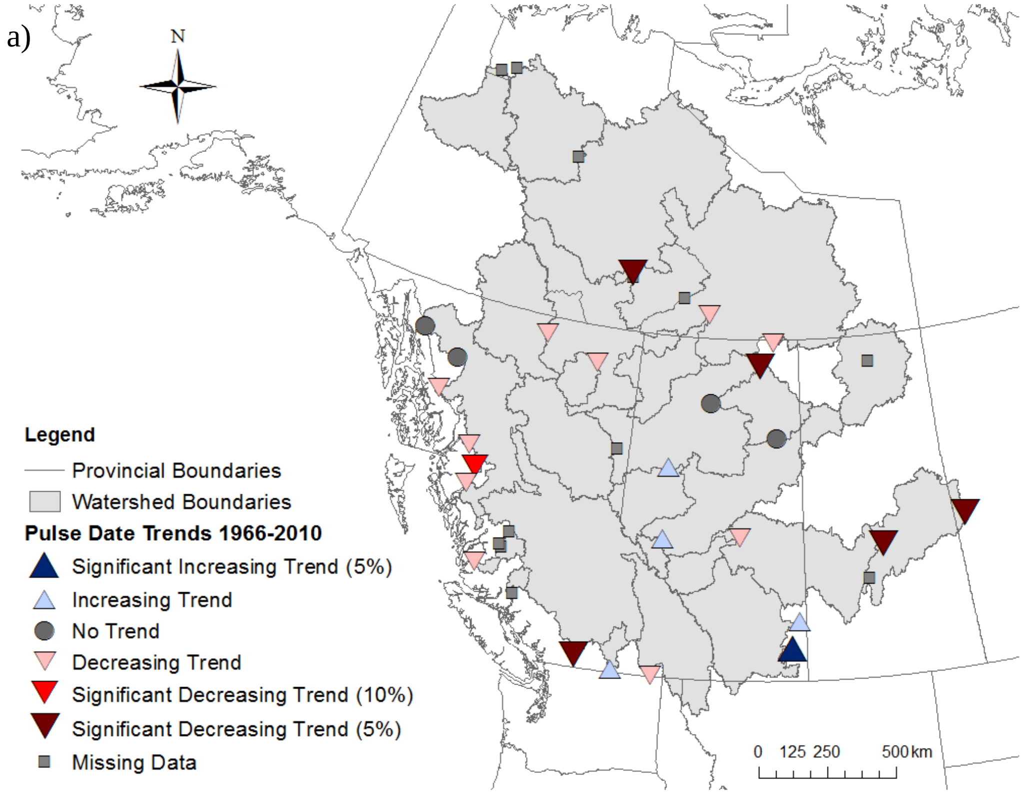


Figure B.4. Same as Figure B.1 but for a) cold season and b) warm season runoff. Note that scale differs from Figure B.2 and Figure B.3.



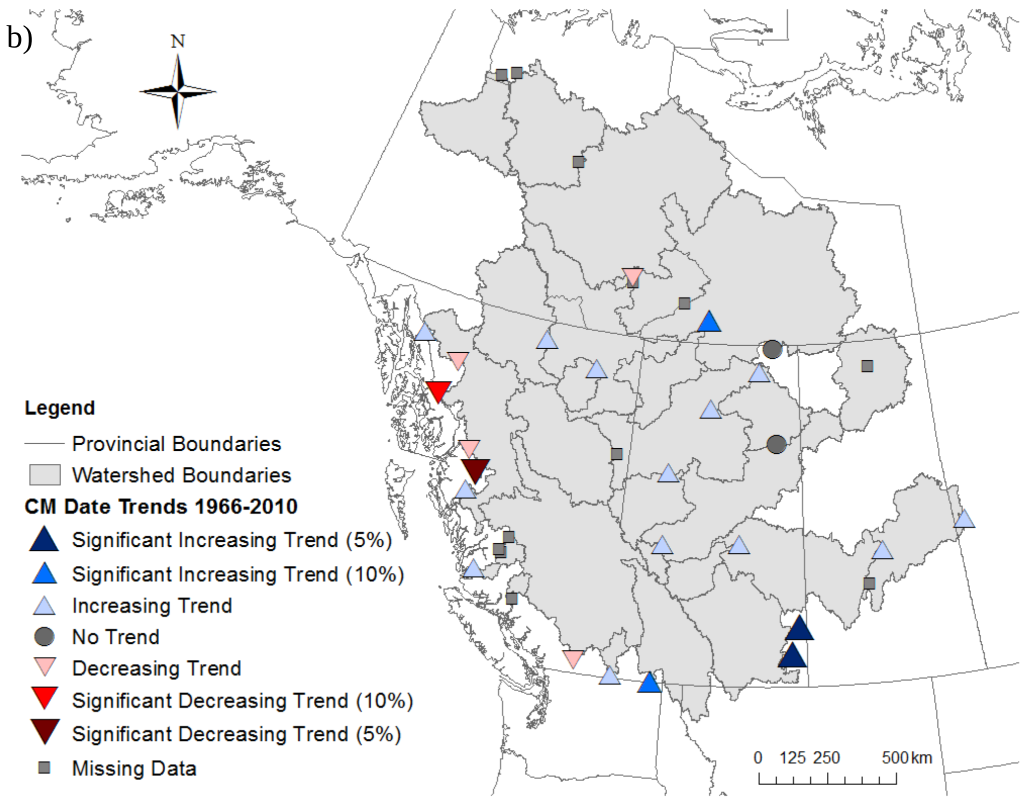


Figure B.5. Map of a) pulse date and b) CM date trends for the 1966-2010 analysis period.

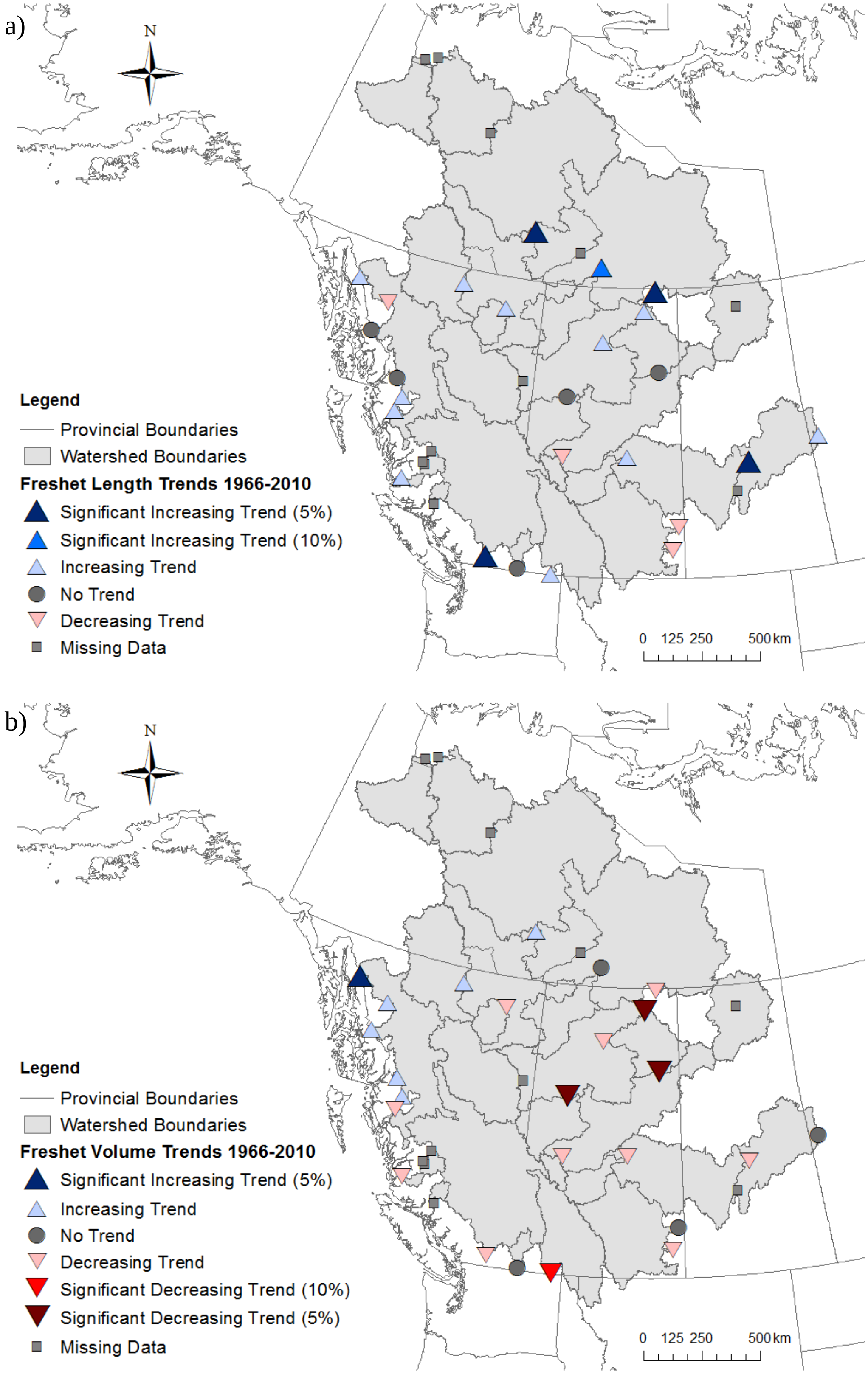


Figure B.6. Map of a) freshet length and b) freshet volume (measured as runoff) trends for the 1966-2010 analysis period.

Appendix C

Principal Component Analysis

Table C.1. Correlation matrix between watersheds, annual runoff, 1976-2010.

WS	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25
1	1	0.24	0.72	0.26	-0.01	0.15	-0.14	-0.16	0.17	0.10	0.56	0.07	0.43	0.08	0.24	0.13	0.71	0.26	0.12	-0.42	-0.05	-0.18	0.14	-0.10	0.21
2	0.24	1	0.34	0.51	0.38	0.55	-0.07	0.41	0.31	0.53	0.41	-0.45	-0.20	-0.21	0.02	-0.09	0.10	0.00	0.32	0.20	0.12	-0.10	-0.11	-0.04	-0.02
3	0.72	0.34	1	0.42	0.11	0.39	0.01	0.02	0.14	0.33	0.76	0.06	0.65	0.05	0.31	0.12	0.43	0.33	0.30	-0.16	0.14	-0.23	0.16	0.00	0.37
4	0.26	0.51	0.42	1	0.21	0.46	-0.08	0.13	0.46	0.34	0.42	-0.25	-0.06	-0.14	-0.05	-0.18	0.23	0.25	0.29	0.09	0.27	-0.24	-0.12	-0.21	0.07
5	-0.01	0.38	0.11	0.21	1	0.55	0.45	0.78	0.27	0.53	0.13	-0.42	-0.29	-0.04	-0.31	-0.03	0.11	0.27	0.60	0.37	0.53	0.25	0.34	0.07	0.01
6	0.15	0.55	0.39	0.46	0.55	1	0.16	0.68	0.27	0.85	0.42	-0.30	-0.07	-0.06	0.03	0.09	0.12	0.24	0.49	0.49	0.54	-0.08	0.32	0.13	0.19
7	-0.14	-0.07	0.01	-0.08	0.45	0.16	1	0.39	0.19	0.22	-0.04	0.04	-0.15	0.14	-0.33	-0.15	0.00	0.52	0.57	0.40	0.56	0.66	0.30	0.43	0.03
8	-0.16	0.41	0.02	0.13	0.78	0.68	0.39	1	0.19	0.81	0.22	-0.50	-0.40	-0.08	-0.26	0.04	-0.15	0.18	0.34	0.40	0.42	0.31	0.39	0.17	0.15
9	0.17	0.31	0.14	0.46	0.27	0.27	0.19	0.19	1	0.23	0.21	0.11	-0.19	-0.01	-0.36	-0.07	0.23	0.03	0.29	0.01	0.15	-0.06	-0.20	-0.23	-0.18
10	0.10	0.53	0.33	0.34	0.53	0.85	0.22	0.81	0.23	1	0.42	-0.33	-0.16	-0.04	-0.08	0.12	0.10	0.18	0.35	0.44	0.47	0.02	0.38	0.16	0.29
11	0.56	0.41	0.76	0.42	0.13	0.42	-0.04	0.22	0.21	0.42	1	-0.01	0.49	0.11	0.16	0.13	0.17	0.13	0.12	-0.22	-0.03	-0.23	0.03	-0.25	0.24
12	0.07	-0.45	0.06	-0.25	-0.42	-0.30	0.04	-0.50	0.11	-0.33	-0.01	1	0.48	0.22	0.09	0.22	0.16	-0.06	-0.08	-0.13	-0.13	-0.11	-0.06	-0.12	0.03
13	0.43	-0.20	0.65	-0.06	-0.29	-0.07	-0.15	-0.40	-0.19	-0.16	0.49	0.48	1	0.35	0.37	0.12	0.11	0.13	-0.01	-0.27	-0.13	-0.33	0.07	-0.08	0.37
14	0.08	-0.21	0.05	-0.14	-0.04	-0.06	0.14	-0.08	-0.01	-0.04	0.11	0.22	0.35	1	-0.42	0.11	-0.07	0.00	0.15	0.01	0.19	-0.05	-0.01	-0.14	-0.13
15	0.24	0.02	0.31	-0.05	-0.31	0.03	-0.33	-0.26	-0.36	-0.08	0.16	0.09	0.37	-0.42	1	0.02	-0.03	-0.05	-0.17	-0.26	-0.27	-0.23	0.04	0.06	0.39
16	0.13	-0.09	0.12	-0.18	-0.03	0.09	-0.15	0.04	-0.07	0.12	0.13	0.22	0.12	0.11	0.02	1	0.02	-0.34	-0.04	-0.01	0.09	-0.31	0.01	-0.17	0.09
17	0.71	0.10	0.43	0.23	0.11	0.12	0.00	-0.15	0.23	0.10	0.17	0.16	0.11	-0.07	-0.03	0.02	1	0.36	0.25	-0.13	0.19	-0.04	0.20	0.03	-0.02
18	0.26	0.00	0.33	0.25	0.27	0.24	0.52	0.18	0.03	0.18	0.13	-0.06	0.13	0.00	-0.05	-0.34	0.36	1	0.47	0.11	0.43	0.31	0.30	0.46	0.09
19	0.12	0.32	0.30	0.29	0.60	0.49	0.57	0.34	0.29	0.35	0.12	-0.08	-0.01	0.15	-0.17	-0.04	0.25	0.47	1	0.63	0.82	0.19	0.30	0.26	0.06
20	-0.42	0.20	-0.16	0.09	0.37	0.49	0.40	0.40	0.01	0.44	-0.22	-0.13	-0.27	0.01	-0.26	-0.01	-0.13	0.11	0.63	1	0.74	0.22	0.33	0.36	0.12
21	-0.05	0.12	0.14	0.27	0.53	0.54	0.56	0.42	0.15	0.47	-0.03	-0.13	-0.13	0.19	-0.27	0.09	0.19	0.43	0.82	0.74	1	0.25	0.40	0.35	0.16
22	-0.18	-0.10	-0.23	-0.24	0.25	-0.08	0.66	0.31	-0.06	0.02	-0.23	-0.11	-0.33	-0.05	-0.23	-0.31	-0.04	0.31	0.19	0.22	0.25	1	0.27	0.55	-0.09
23	0.14	-0.11	0.16	-0.12	0.34	0.32	0.30	0.39	-0.20	0.38	0.03	-0.06	0.07	-0.01	0.04	0.01	0.20	0.30	0.30	0.33	0.40	0.27	1	0.44	0.63
24	-0.10	-0.04	0.00	-0.21	0.07	0.13	0.43	0.17	-0.23	0.16	-0.25	-0.12	-0.08	-0.14	0.06	-0.17	0.03	0.46	0.26	0.36	0.35	0.55	0.44	1	0.17
25	0.21	-0.02	0.37	0.07	0.01	0.19	0.03	0.15	-0.18	0.29	0.24	0.03	0.37	-0.13	0.39	0.09	-0.02	0.09	0.06	0.12	0.16	-0.09	0.63	0.17	1

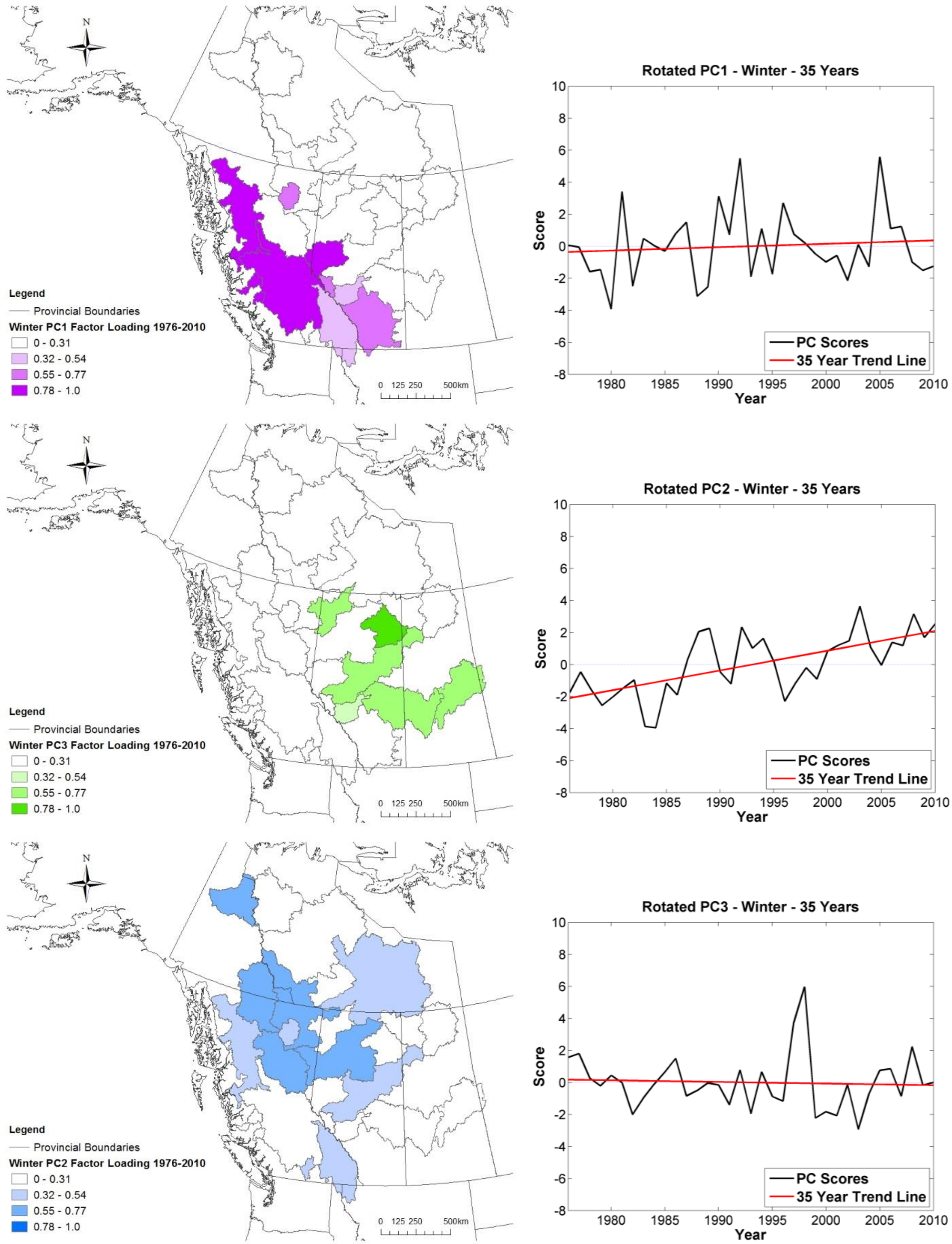


Figure C.1. Factor loading maps for the first three rotated PCs for winter runoff (left hand side) and corresponding score plots for each PC (right hand side).

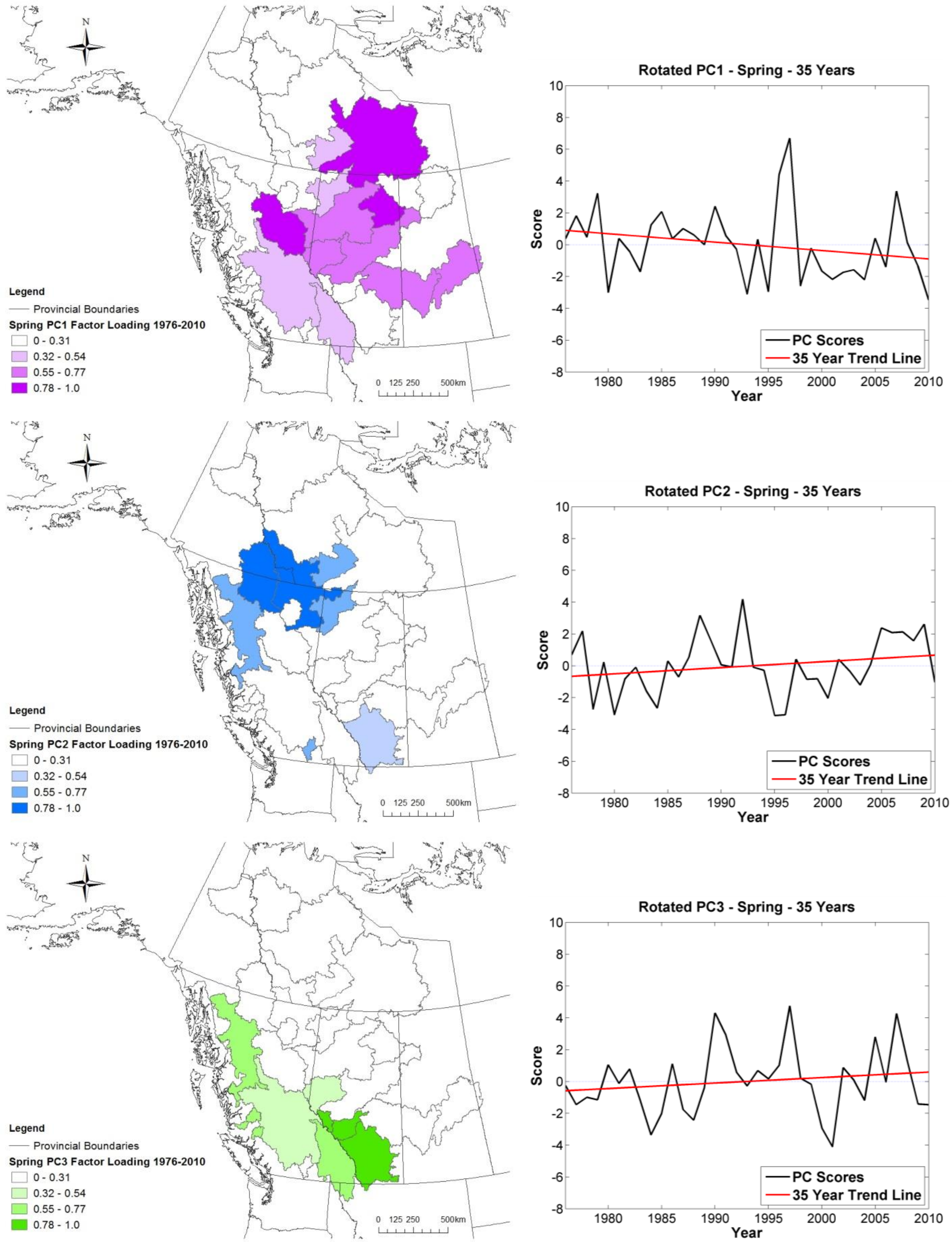


Figure C.2. Factor loading maps for the first three rotated PCs for spring runoff (left hand side) and corresponding score plots for each PC (right hand side).

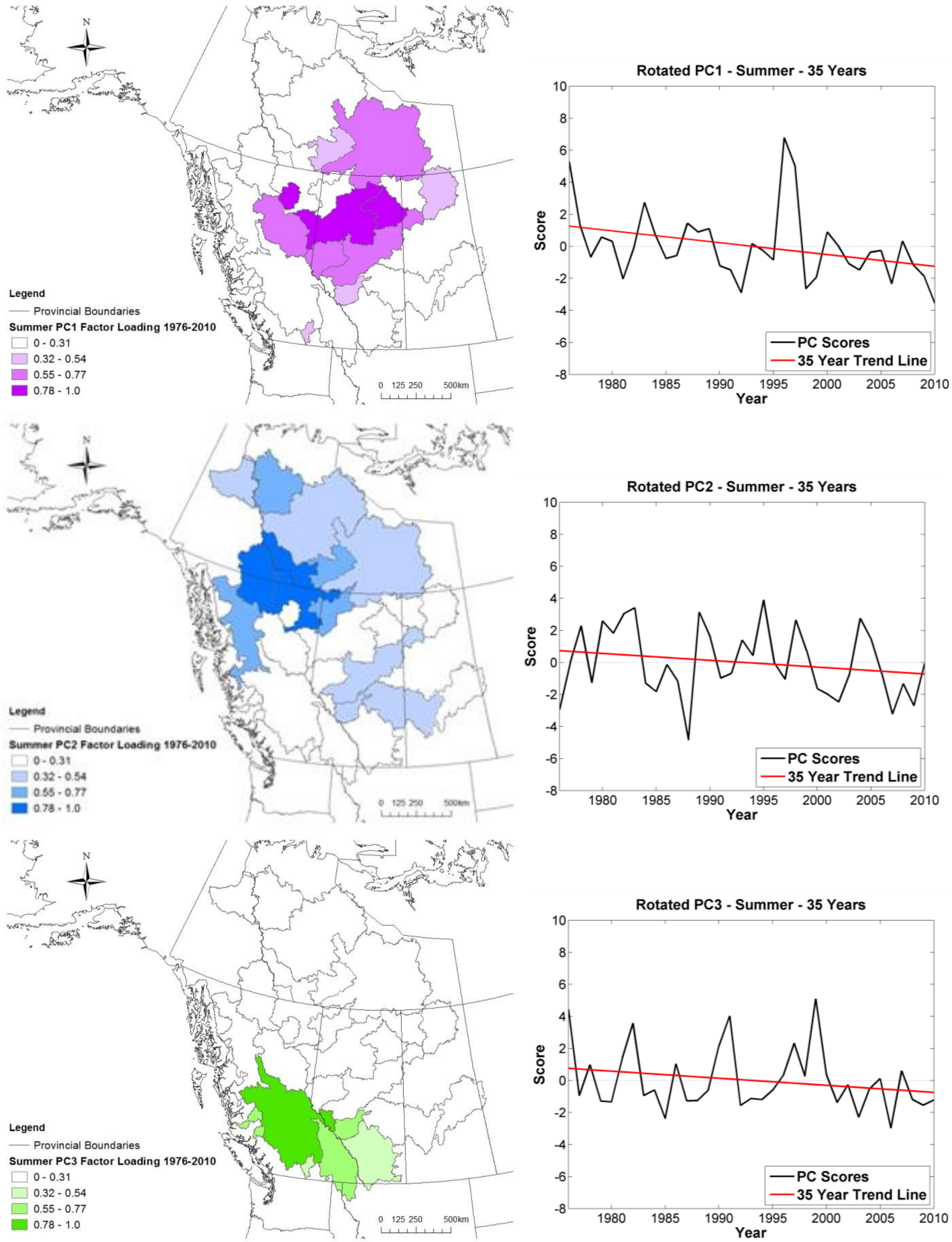


Figure C.3. Factor loading maps for the first three rotated PCs for summer runoff (left hand side) and corresponding score plots for each PC (right hand side).

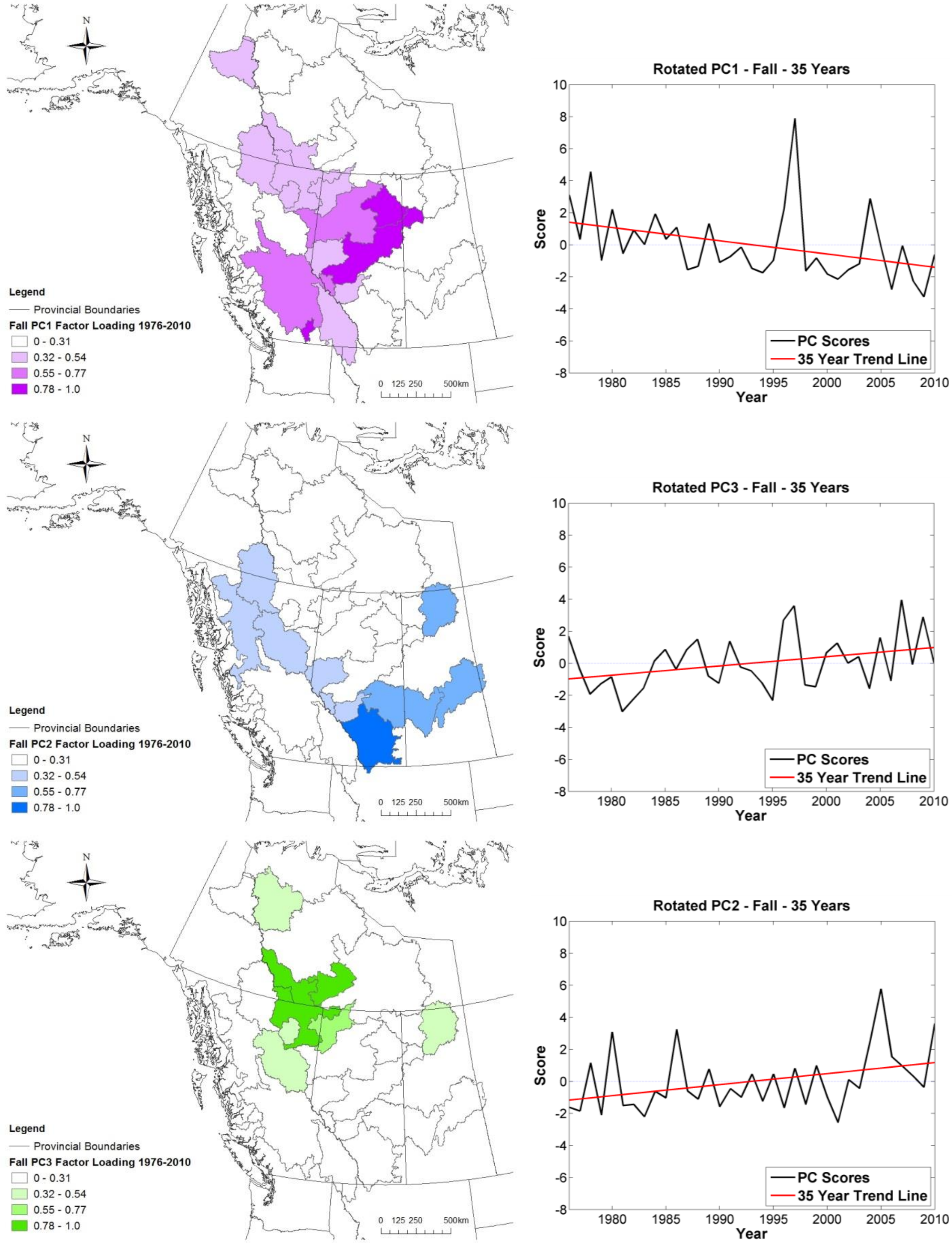


Figure C.4. Factor loading maps for the first three rotated PCs for fall runoff (left hand side) and corresponding score plots for each PC (right hand side).

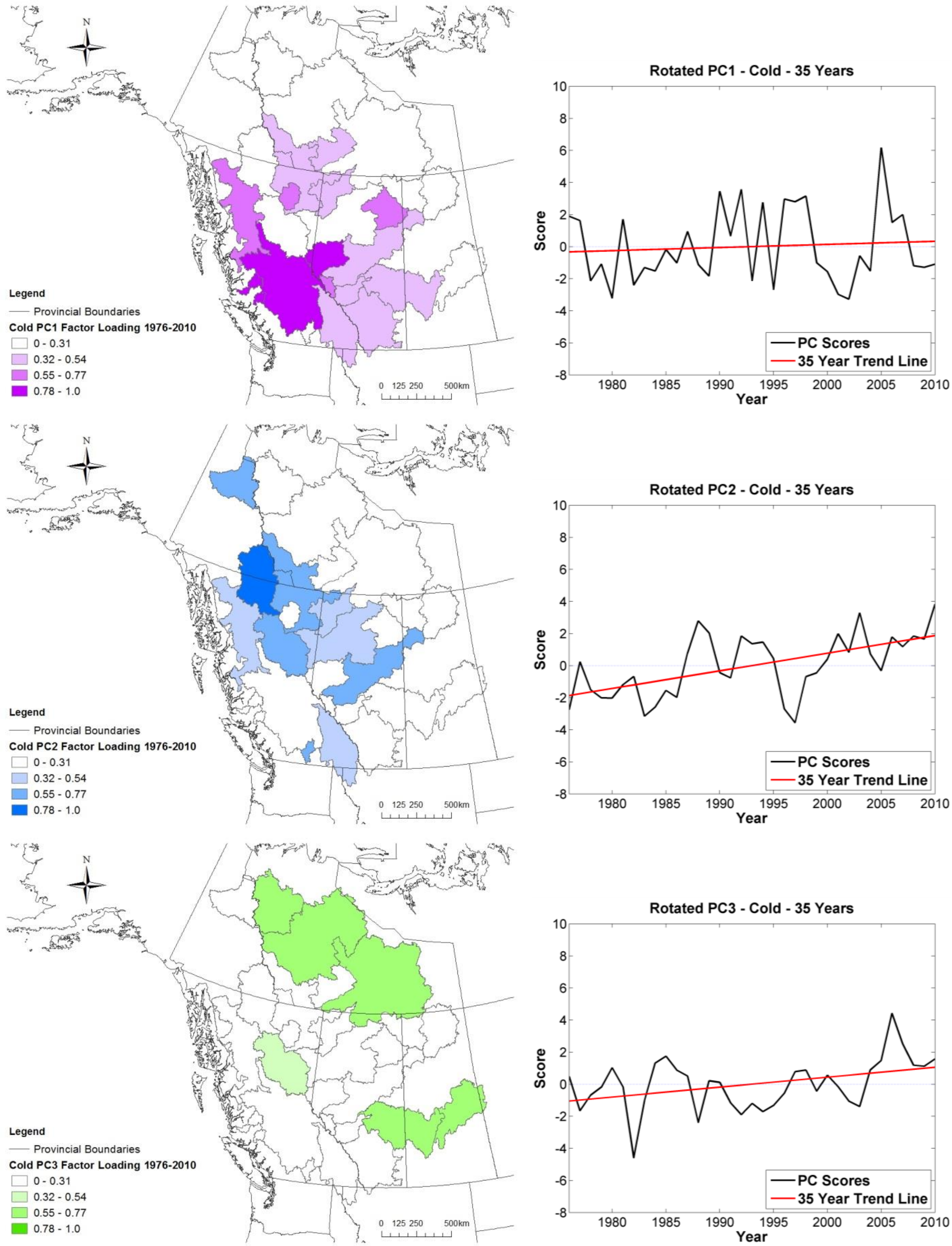


Figure C.5. Factor loading maps for the first three rotated PCs for cold season runoff (left hand side) and corresponding score plots for each PC (right hand side).

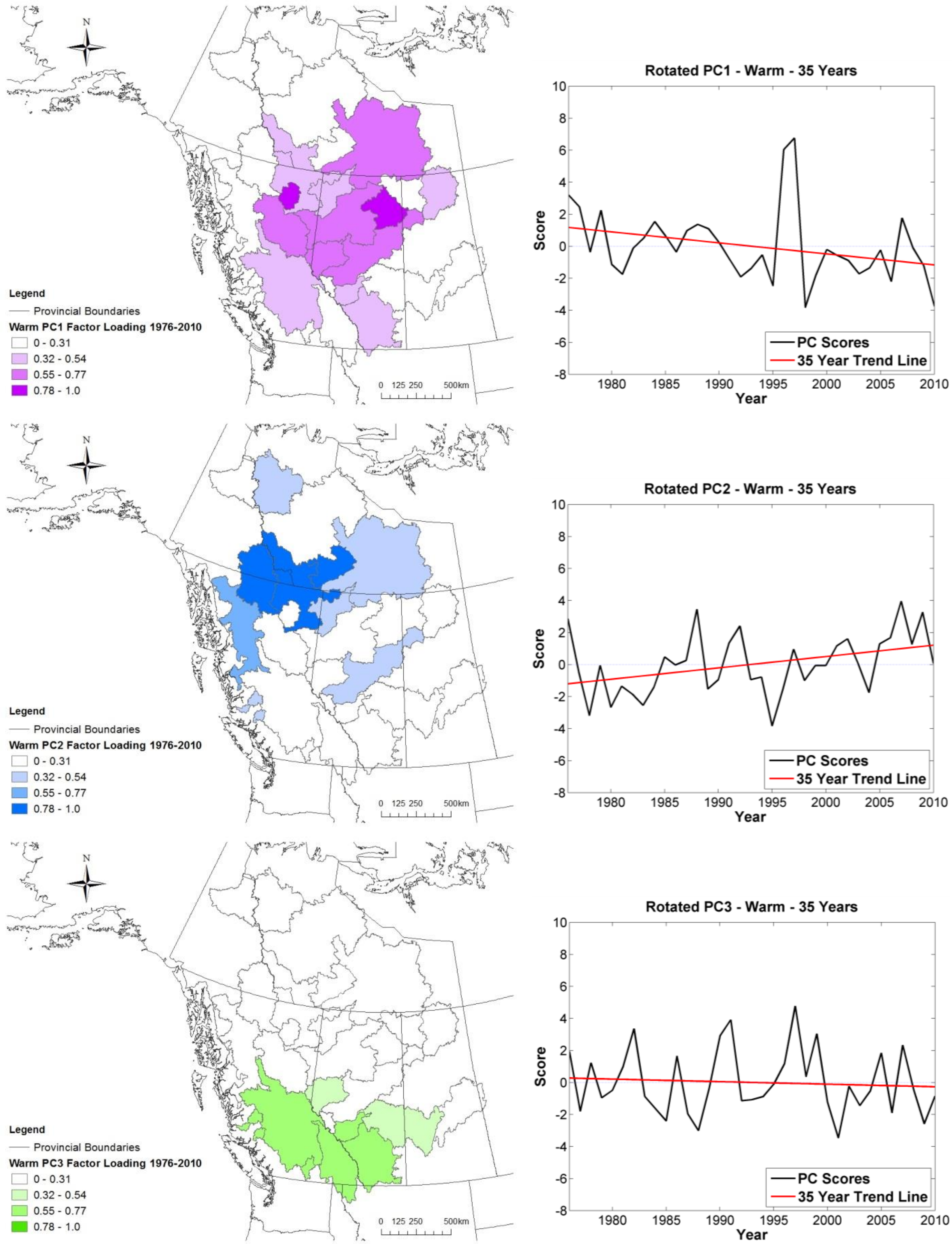


Figure C.6. Factor loading maps for the first three rotated PCs for warm season runoff (left hand side) and corresponding score plots for each PC (right hand side).

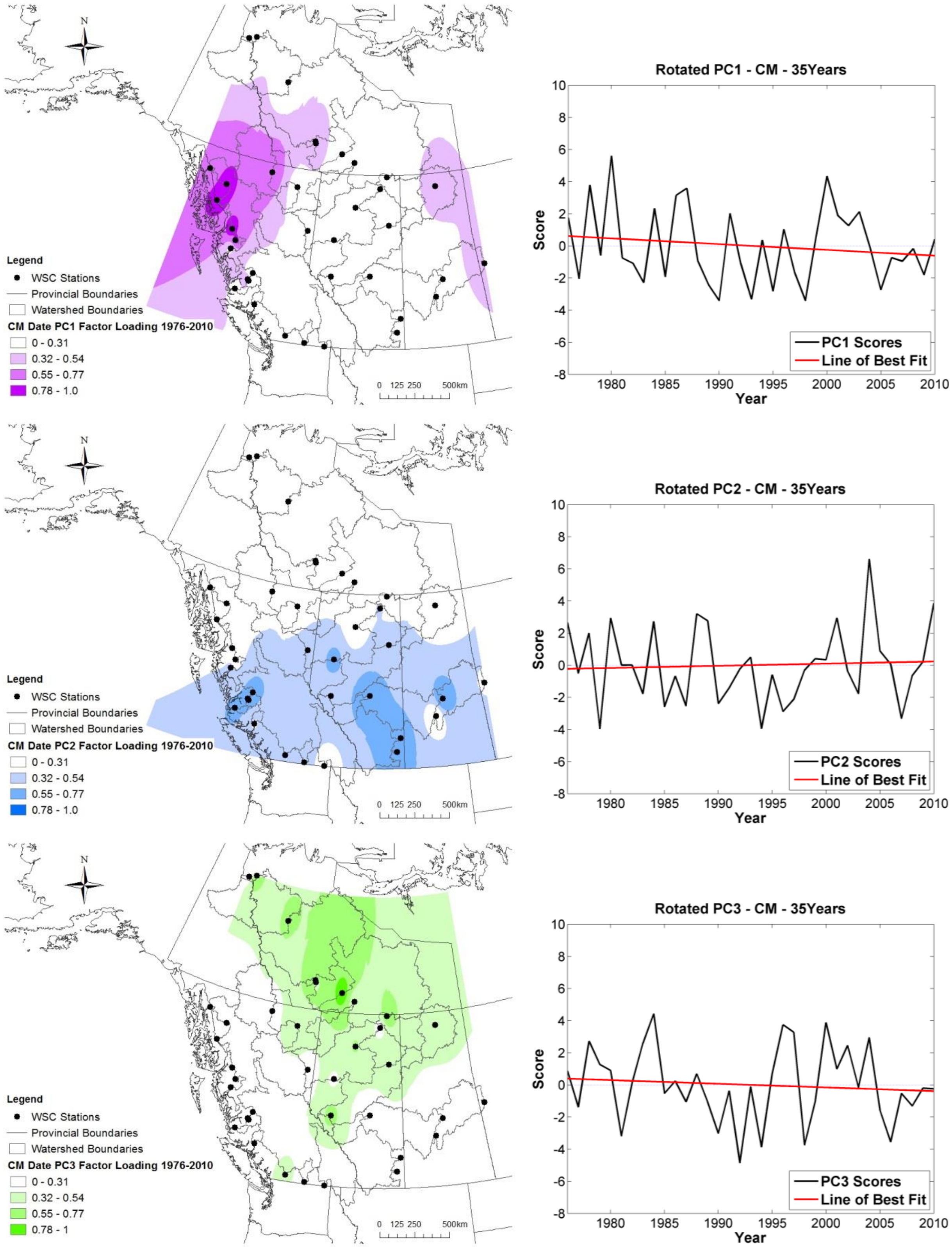


Figure C.7. Factor loading maps for the first three rotated PCs for CM date (left hand side) and corresponding score plots for each PC (right hand side).

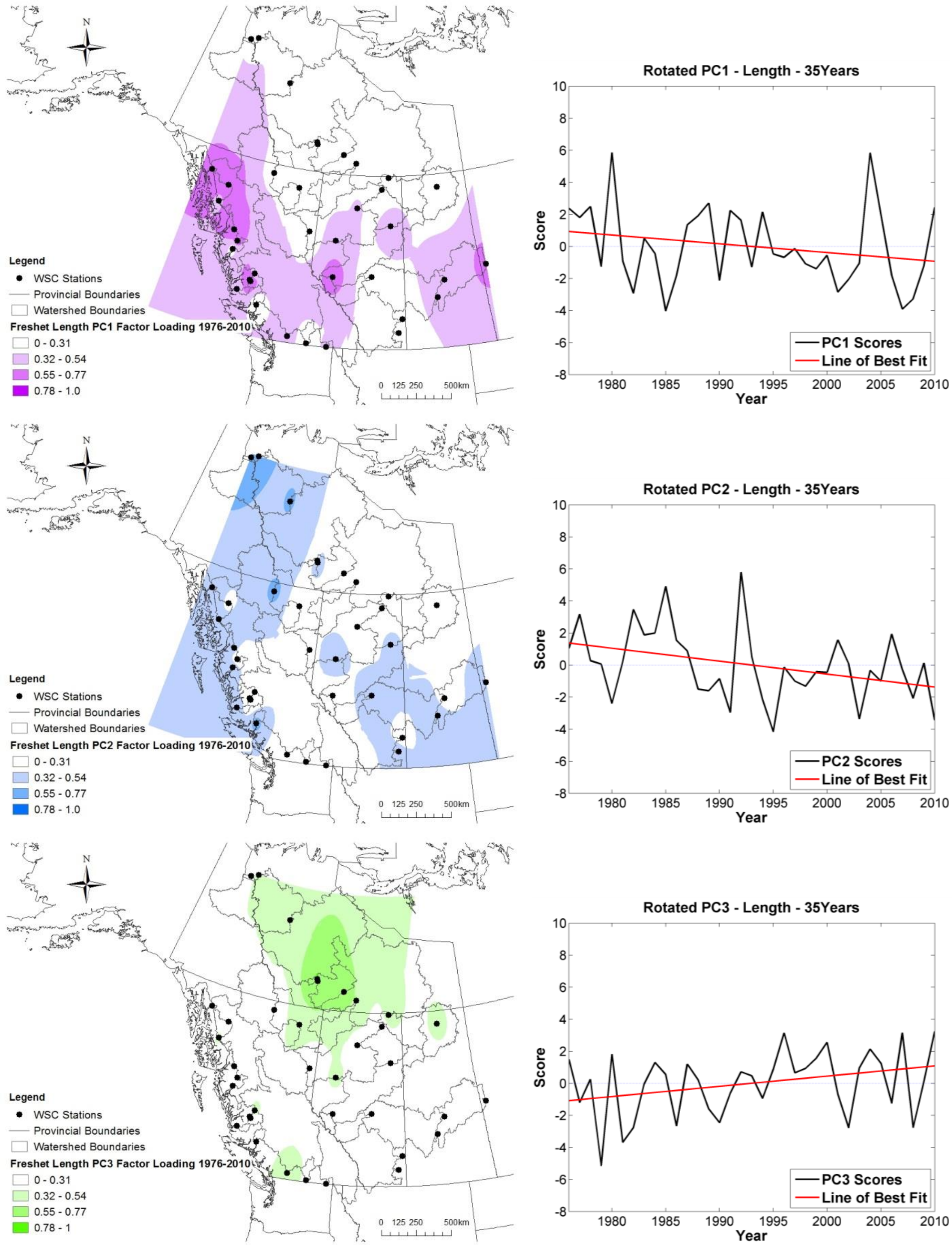


Figure C.8. Factor loading maps for the first three rotated PCs for freshet length (left hand side) and corresponding score plots for each PC (right hand side).

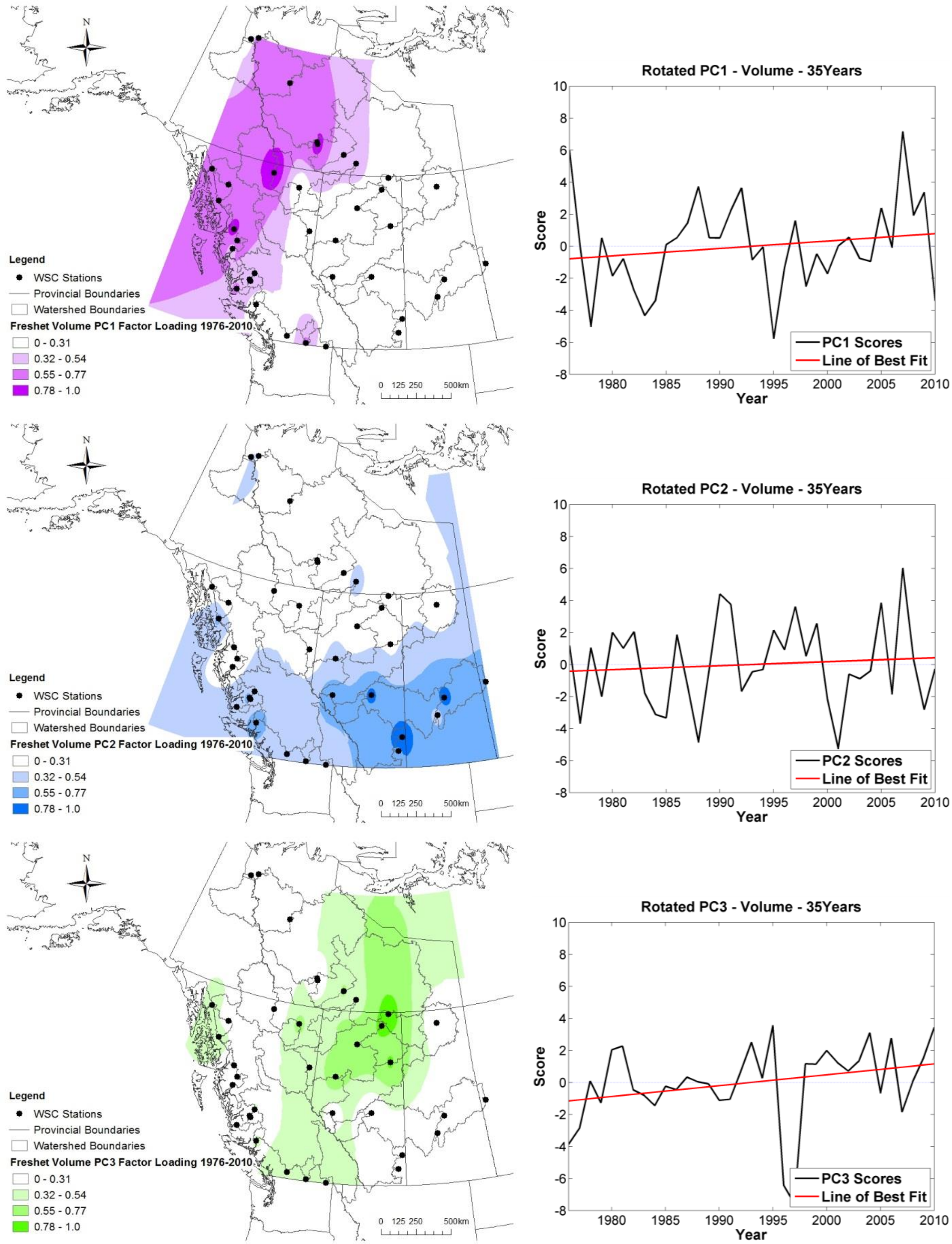


Figure C.9. Factor loading maps for the first three rotated PCs for freshet volume runoff (left hand side) and corresponding score plots for each PC (right hand side).

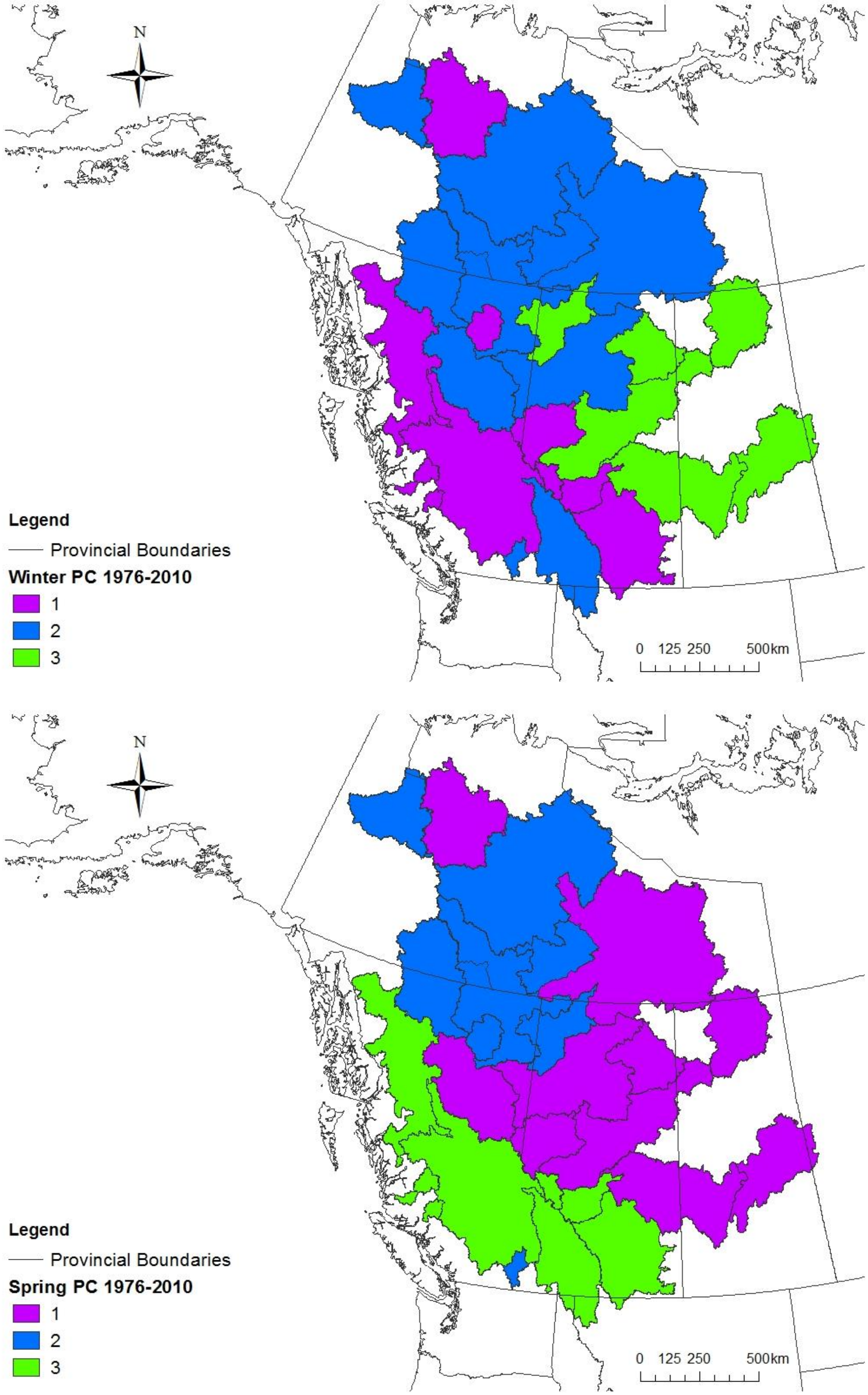


Figure C.10. Factor loading to which each basin loaded highest in a) winter and b) spring runoff PCA.

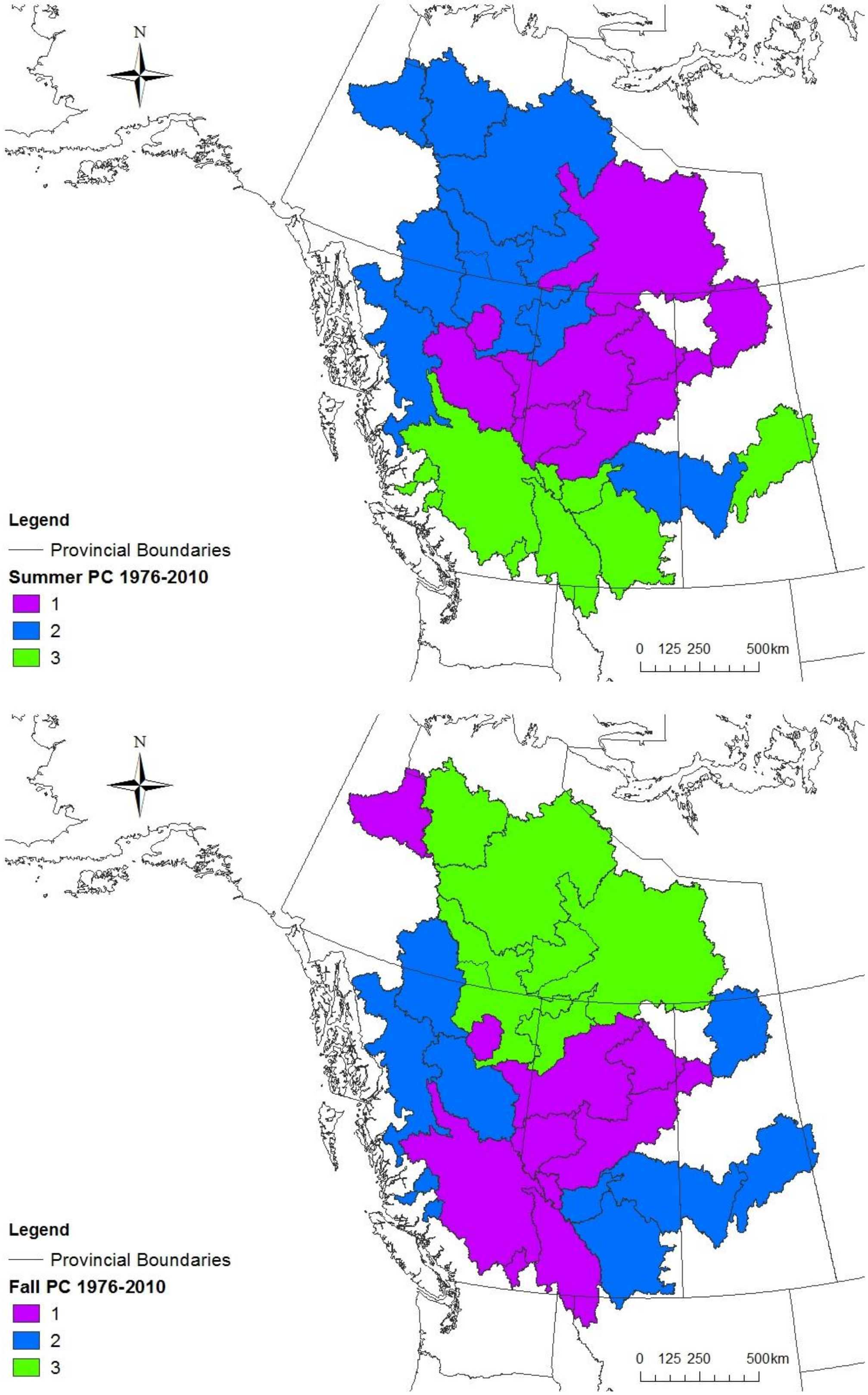


Figure C.11. Factor loading to which each basin loaded highest in a) summer and b) fall runoff PCA.

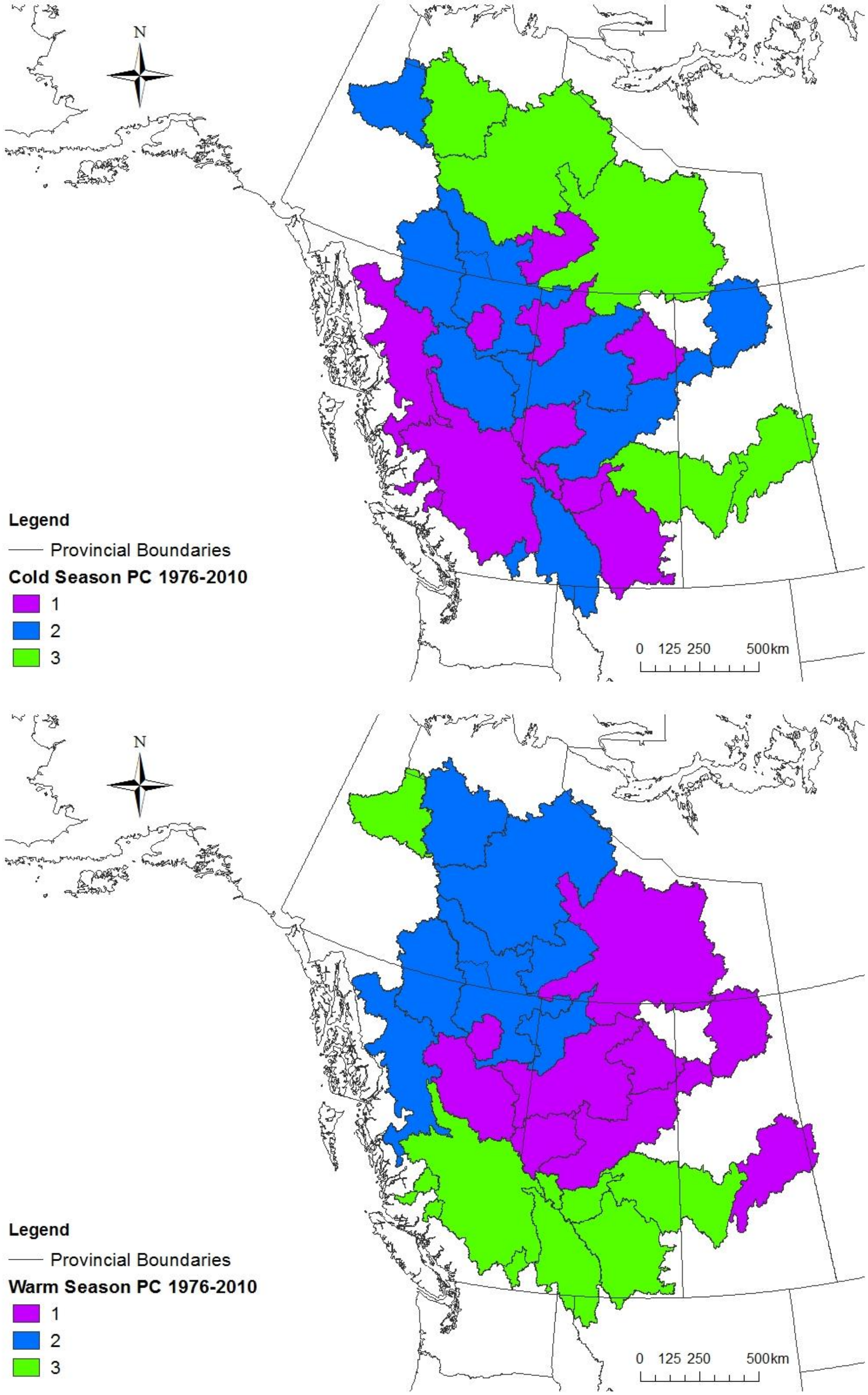


Figure C.12. Factor loading to which each basin loaded highest in a) cold season and b) warm season runoff PCA.

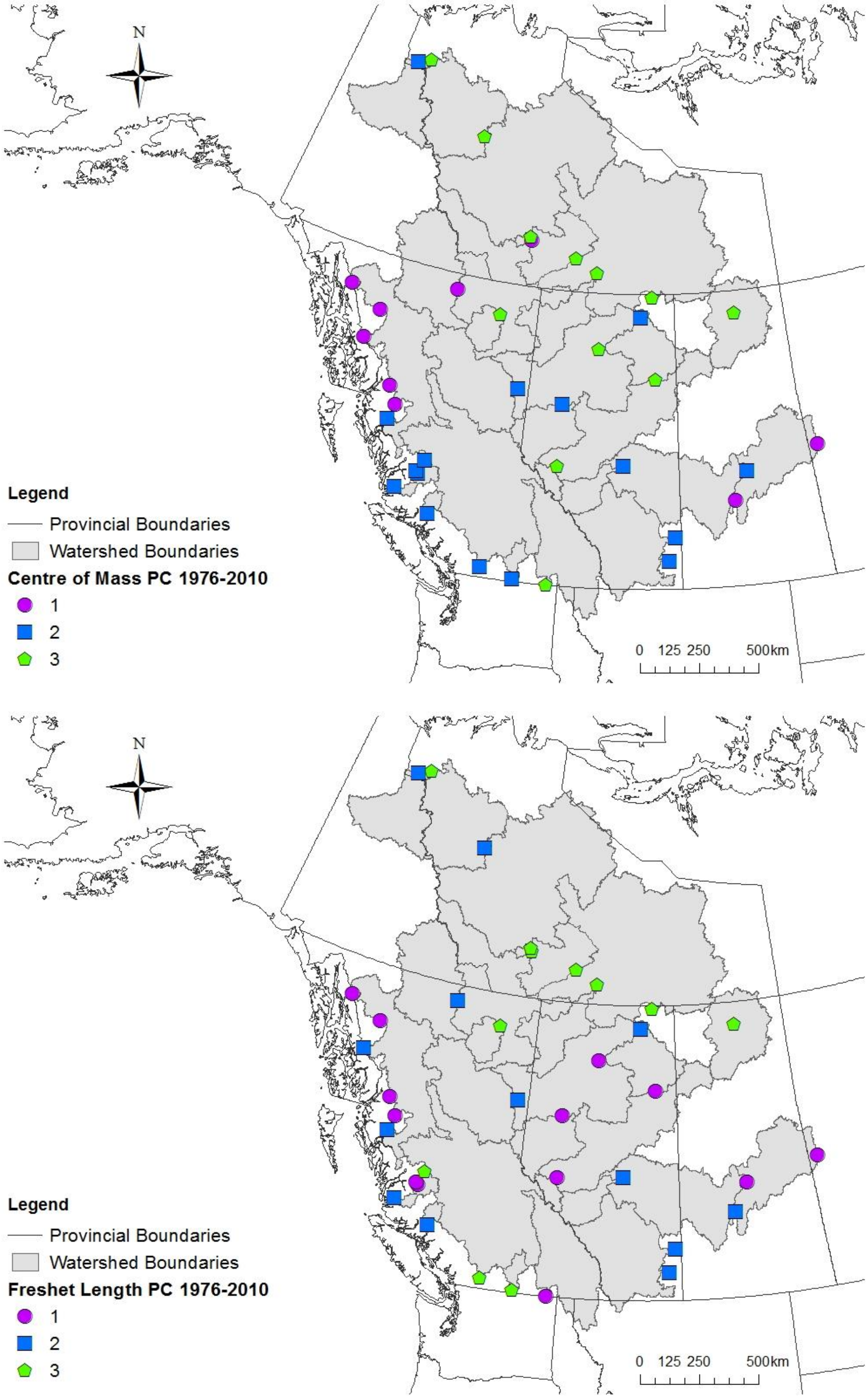


Figure C.13. Factor loading to which each station loaded highest in a) CM date and b) freshet length PCA.

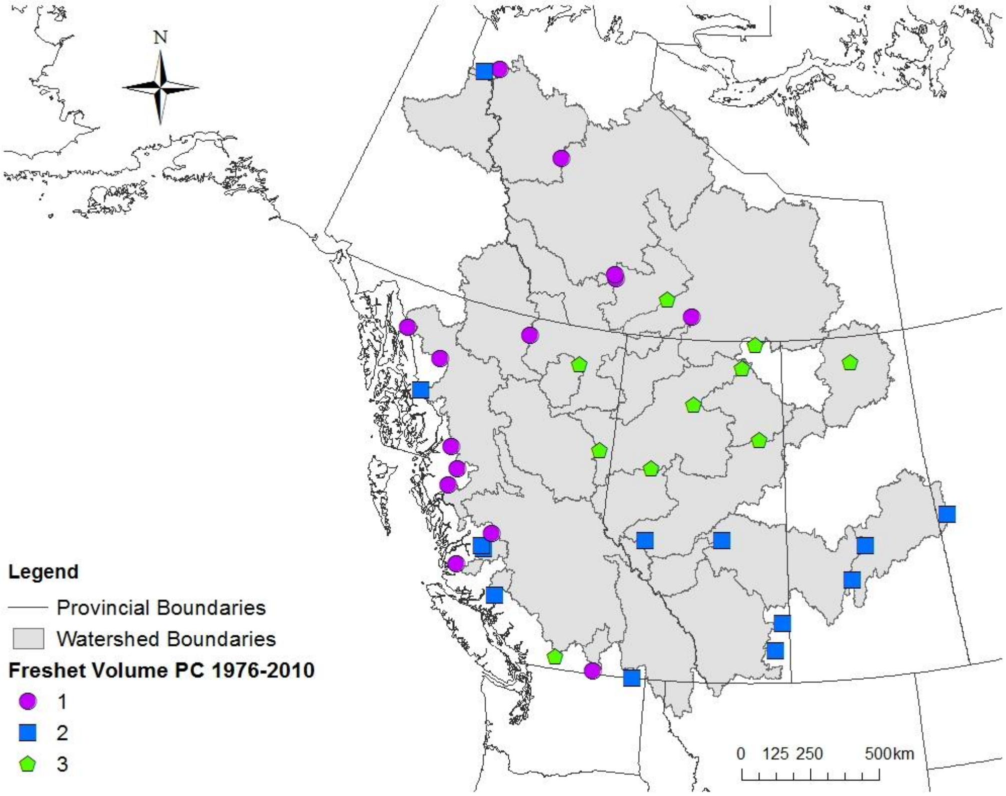


Figure C.14. Factor loading to which each station loaded highest freshet volume PCA.

Table C.2. PCA Region Slopes in anomaly/35-years

		Northern Region	Middle Region	Southern Region
Annual (mm/35yrs)	**Slope**	3.2	-2.9	0.1
	Trend	↑↑	↓↓	↑
Winter (mm/35yrs)	**Slope**	-1.0	4.0	0
	Trend	↓	↑↑	None
Spring (mm/35yrs)	**Slope**	1.0	-3.0	1.0
	Trend	↑	↓↓	↑
Summer (mm/35yrs)	**Slope**	-2.0	-3.0	-1.0
	Trend	↓	↓	↓
Fall (mm/35yrs)	**Slope**	2.0	-3.0	2.0
	Trend	↑↑	↓↓	↑↑
Cold Season (mm/35yrs)	**Slope**	1.9	4.0	0.4
	Trend	↑	↑↑	↑
Warm Season (mm/35yrs)	**Slope**	2.9	-3.0	-0.5
	Trend	↑↑	↓↓	↓

↑ or ↓: Non-significant trend
↑↑ or **↓↓**: Trend significant at 10% significance level
↑↑ or ↓↓: Trend significant at 10% significance level

Appendix D

Climatic Relationships

Table D.1. Maximum temperature trend results for the 1976-2010 (35-year) period.

Watershed	Annual[+] (mm/yr)		Winter (mm/yr)		Spring[+] (mm/yr)		Summer[++] (mm/yr)		Fall (mm/yr)		Cold Season (mm/yr)		Warm Season (mm/yr)	
	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend
Upper Liard	0.3	↑	-2.2	↓	1.5	↑↑	0.7	↑	-0.9	↓	0	None	0.8	↑
Fort Nelson	0.7	↑	-0.3	↓	0.3	↑	0.7	↑	0.3	↑	1.1	↑	0.3	↑
Lower Liard	0.8	↑	0.2	↑	1	↑↑	0.7	↑	0	None	0.8	↑	0.4	↑
Upper Peace	0.4	↑	-0.9	↓	0.4	↑	0.8	↑	-0.6	↓	0.6	↑	0.5	↑
Smoky River	0.3	↑	0	None	-0.5	↓	1.2	↑↑	-0.7	↓	-0.1	↓	0.3	↑
Lower Peace	0.7	↑	0.4	↑	-0.2	↓	1.1	↑↑	0.2	↑	0.7	↑	0.3	↑
Upper Athabasca	0.8	↑	1.2	↑	-0.1	↓	1.5	↑↑	-0.4	↓	0.9	↑	0.5	↑
Lower Athabasca	0.3	↑	0.3	↑	-0.7	↓	1.4	↑↑	-0.1	↓	0.7	↑	0.3	↑
East Lake Athabasca	1.3	↑↑	1.4	↑	1.1	↑	1.5	↑↑	2.3	↑	1.8	↑	1.2	↑↑
West Lake Athabasca	0.9	↑	0.7	↑	-0.3	↓	1.2	↑↑	1.6	↑	1.2	↑	0.3	↑
Hay	0.7	↑	0	None	0.3	↑	0.6	↑	1.5	↑	1.1	↑	0.2	↑
Great Slave	1.5	↑	1.6	↑	0.7	↑	0.7	↑	3.1	↑↑	2.3	↑↑	0.4	↑
Upper Mackenzie	1.4	↑	0.8	↑	1	↑	0.4	↑	2.6	↑↑	2.4	↑↑	0.4	↑
Mid Mackenzie	1.3	↑	0.8	↑	1.1	↑↑	0.3	↑	1.9	↑↑	1.9	↑	0.3	↑
Lower Mackenzie	1.1	↑	0.2	↑	2.8	↑↑	0.1	↑	1.5	↑	1.8	↑	0.8	↑↑
Peel	1.1	↑	0.2	↑	2.8	↑↑	-0.2	↓	1.5	↑	1.9	↑↑	1	↑
North Pacific	-0.1	↓	-1.3	↓	0.1	↑	-0.5	↓	-1.1	↓	-0.7	↓	-0.2	↓
South Pacific	0.2	↑	-0.2	↓	0.4	↑	0.9	↑	-1.2	↓	-0.1	↓	0.6	↑
Fraser	0.8	↑	1.1	↑	0.4	↑	1.4	↑↑	-0.5	↓	0.8	↑	0.9	↑
Okanagan	1.4	↑↑	1.3	↑↑	0.9	↑	2.8	↑↑	0.1	↑	1.2	↑↑	1.8	↑↑
Columbia	1.2	↑↑	0.9	↑	0.4	↑	2.6	↑↑	0.3	↑	1	↑↑	1.3	↑↑
Upper North Saskatchewan	0.7	↑	1.1	↑	-0.6	↓	1.6	↑↑	-0.2	↓	1	↑	0.3	↑
Lower North Saskatchewan	-0.1	↓	0	None	-1.4	↓	1.1	↑	0.3	↑	0.4	↑	-0.2	↓
Upper South Saskatchewan	0.3	↑	0.8	↑	-1.3	↓	1.2	↑↑	-0.2	↓	0.8	↑	-0.2	↓
Lower South Saskatchewan	0	None	0.4	↑	-1.4	↓	0.7	↑	1.1	↑	0.9	↑	-0.3	↓

Table D.1(con't). Maximum temperature trend results for the 1976-2010 (35-year) period.

Station	Annual+		Winter		Spring+		Summer++		Fall		Cold Season		Warm Season	
	#	%	#	%	#	%	#	%	#	%	#	%	#	%
# ↓	2	**8%**	5	**20%**	9	**36%**	2	**8%**	10	**40%**	3	**12%**	4	**16%**
# **↓↓**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**
# ↑	19	**76%**	16	**64%**	11	**44%**	12	**48%**	11	**44%**	16	**64%**	17	**68%**
# **↑↑**	3	**12%**	1	**4%**	5	**20%**	11	**44%**	3	**12%**	5	**20%**	4	**16%**
# No trend	1	**4%**	3	**12%**	0	**0%**	0	**0%**	1	**4%**	1	**4%**	0	**0%**
# Missing	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**

+ Variable is field significant at 10% significance level.

++ Variable is field significant at 5% significance level.

↑ or ↓: Non-significant trend

↑↑ or **↓↓**: Trend significant at 10% significance level

↑↑ or **↓↓**: Trend significant at 10% significance level

Table D.2. Minimum temperature trend results for the 1976-2010 (35-year) period.

Watershed	Annual[+] (mm/yr)		Winter (mm/yr)		Spring (mm/yr)		Summer[++] (mm/yr)		Fall (mm/yr)		Cold Season (mm/yr)		Warm Season[+] (mm/yr)	
	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend	Slope	Trend
Upper Liard	0.5	↑	-1.7	↓	0.5	↑	1.1	↑↑	-0.2	↓	0.5	↑	0.6	↑↑
Fort Nelson	0.4	↑	-0.8	↓	-0.3	↓	0.2	↑	0.3	↑	0.9	↑	-0.4	↓
Lower Liard	0.4	↑	-0.4	↓	0.1	↑	0.4	↑	0.4	↑	0.7	↑	0.1	↑
Upper Peace	0.1	↑	-0.7	↓	-0.2	↓	0.2	↑	-0.4	↓	0.2	↑	-0.2	↓
Smoky River	-0.9	↓	-1.1	↓	-1.7	↓↓	-0.9	↓	-1.1	↓	-0.3	↓	-1.3	↓↓
Lower Peace	0.6	↑	0.9	↑	-0.7	↓	-0.1	↓	1.5	↑	1.2	↑	-0.5	↓
Upper Athabasca	0.9	↑	2.3	↑	-0.5	↓	0.5	↑	1.2	↑	2.2	↑	-0.2	↓
Lower Athabasca	0.4	↑	1.1	↑	-1	↓↓	-0.1	↓	1.1	↑	1.5	↑	-0.4	↓
East Lake Athabasca	0.7	↑	2	↑	-1.2	↓	0.2	↑	2.2	↑	1.7	↑	-0.3	↓
West Lake Athabasca	1.2	↑↑	2.1	↑	-0.8	↓	0.6	↑↑	3.2	↑↑	2.3	↑	0.1	↑
Hay	0.8	↑	0.9	↑	-0.8	↓	-0.1	↓	2.2	↑↑	2	↑	-0.4	↓
Great Slave	1.7	↑↑	2.5	↑	0.5	↑	1.3	↑↑	3.8	↑↑	3.1	↑↑	0.7	↑↑
Upper Mackenzie	1.1	↑	1.3	↑	-0.4	↓	0.5	↑	2.3	↑↑	2.4	↑	0.1	↑
Mid Mackenzie	1	↑	0.7	↑	0.4	↑	0.4	↑	1.7	↑	1.5	↑	0	None
Lower Mackenzie	1.4	↑↑	0.5	↑	2.8	↑↑	0.8	↑↑	1.7	↑	2.4	↑	1.1	↑↑
Peel	1.6	↑	0.8	↑	2.6	↑↑	0.3	↑	2.3	↑	2.2	↑	0.9	↑
North Pacific	1.2	↑↑	0.7	↑	0.8	↑	1.1	↑↑	0.5	↑	1.3	↑	0.9	↑
South Pacific	1.5	↑↑	0.8	↑	1	↑	1.5	↑↑	0.9	↑	1.5	↑↑	1.3	↑↑
Fraser	0.7	↑	1	↑	0.2	↑	0.6	↑	0.5	↑	0.9	↑	0.3	↑
Okanagan	1.4	↑↑	1.4	↑	0.8	↑	1.8	↑↑	1.2	↑↑	1.2	↑↑	1.5	↑↑
Columbia	0.9	↑↑	1.5	↑	-0.1	↓	0.8	↑↑	1.1	↑	1.4	↑↑	0.4	↑
Upper North Saskatchewan	1.2	↑↑	2.9	↑↑	-0.6	↓	0.1	↑	1.7	↑	2.5	↑↑	-0.3	↓
Lower North Saskatchewan	0.3	↑	0.4	↑	-1	↓	0.4	↑	1.2	↑	0.9	↑	-0.1	↓
Upper South Saskatchewan	0.6	↑	1.6	↑	-0.9	↓	0.1	↑	1.1	↑	1.4	↑	-0.2	↓
Lower South Saskatchewan	0.9	↑	1.3	↑	-0.8	↓	1.5	↑↑	2.6	↑↑	1.8	↑	0.5	↑

Table D.2(con't). Minimum temperature trend results for the 1976-2010 (35-year) period.

Station	Annual[+]		Winter		Spring		Summer[++]		Fall		Cold Season		Warm Season[+]	
	#	**%**	**#**	**%**	**#**	**%**	**#**	**%**	**#**	**%**	**#**	**%**	**#**	**%**
# ↓	10	**40%**	5	**20%**	13	**52%**	4	**16%**	3	**12%**	1	**4%**	10	**40%**
# **↓↓**	1	**4%**	0	**0%**	2	**8%**	0	**0%**	0	**0%**	0	**0%**	1	**4%**
# ↑	8	**32%**	19	**76%**	8	**32%**	12	**48%**	16	**64%**	19	**76%**	8	**32%**
# **↑↑**	5	**20%**	1	**4%**	2	**8%**	9	**36%**	6	**24%**	5	**20%**	5	**20%**
# No trend	1	**4%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	1	**4%**
# Missing	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**	0	**0%**

[+] Variable is field significant at 10% significance level.

[++] Variable is field significant at 5% significance level.

↑ or ↓: Non-significant trend

↑↑ or **↓↓**: Trend significant at 10% significance level

↑↑ or ↓↓: Trend significant at 10% significance level

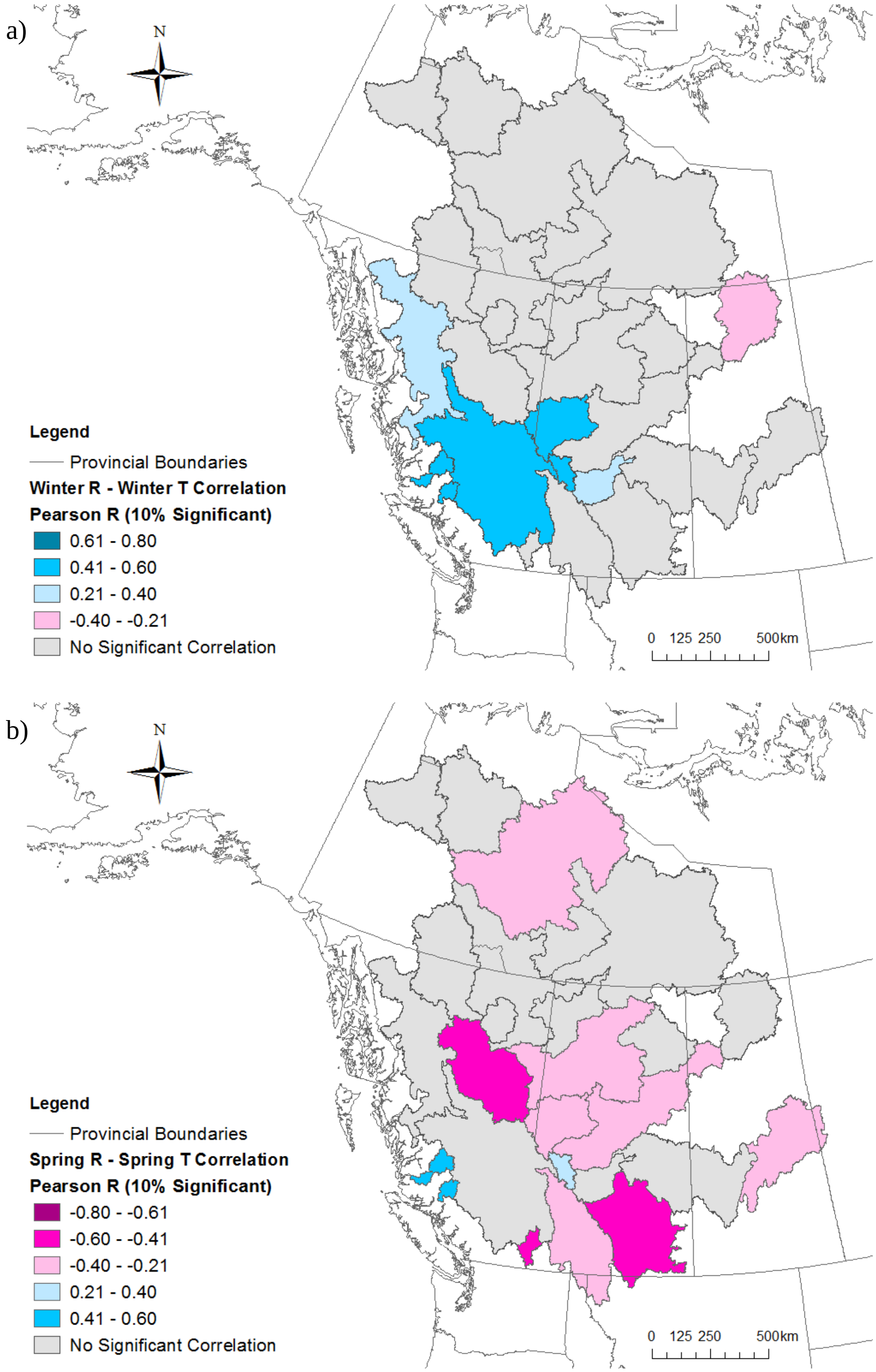
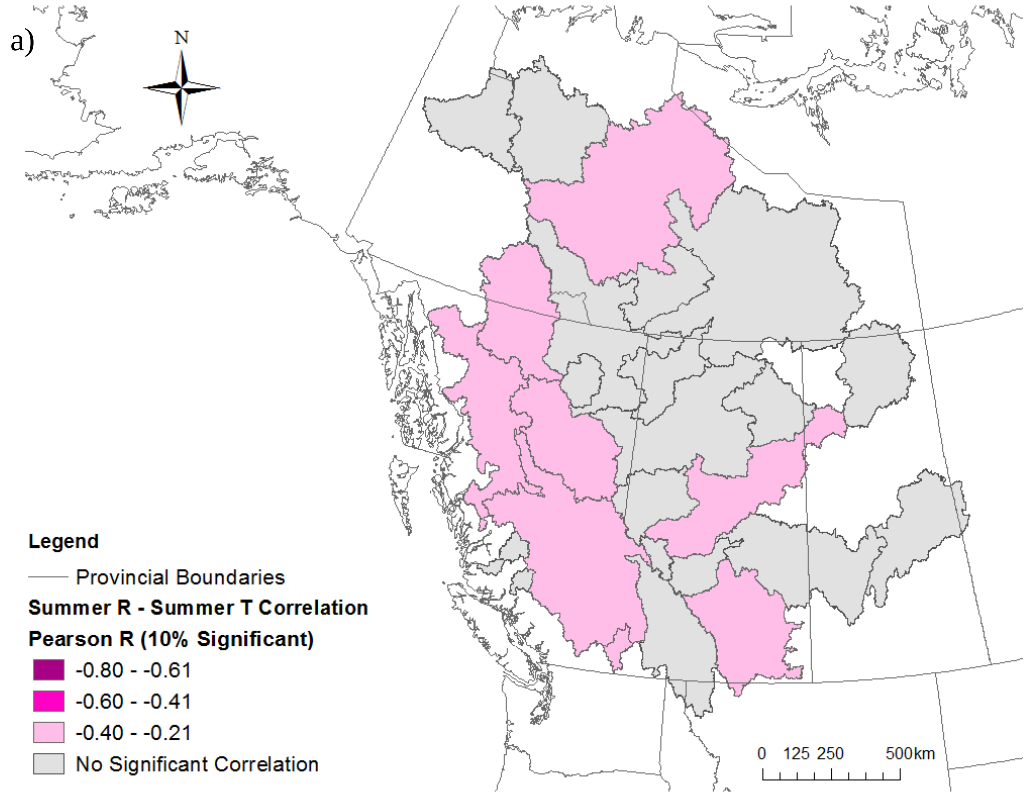


Figure D.1. Basins that exhibited significant correlations ($p < 0.1$) between a) winter and b) spring runoff and maximum temperature of the same season, respectively.



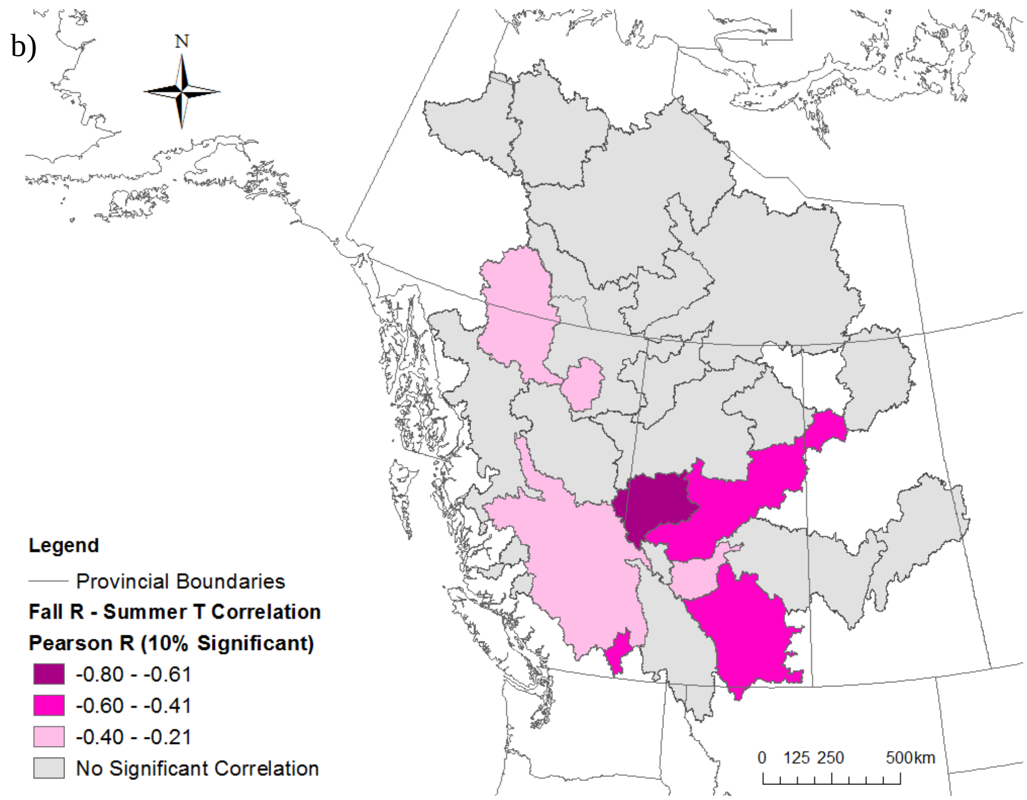


Figure D.2. Basins that exhibited significant correlations ($p < 0.1$) between a) summer and b) fall runoff and maximum summer temperature, respectively.

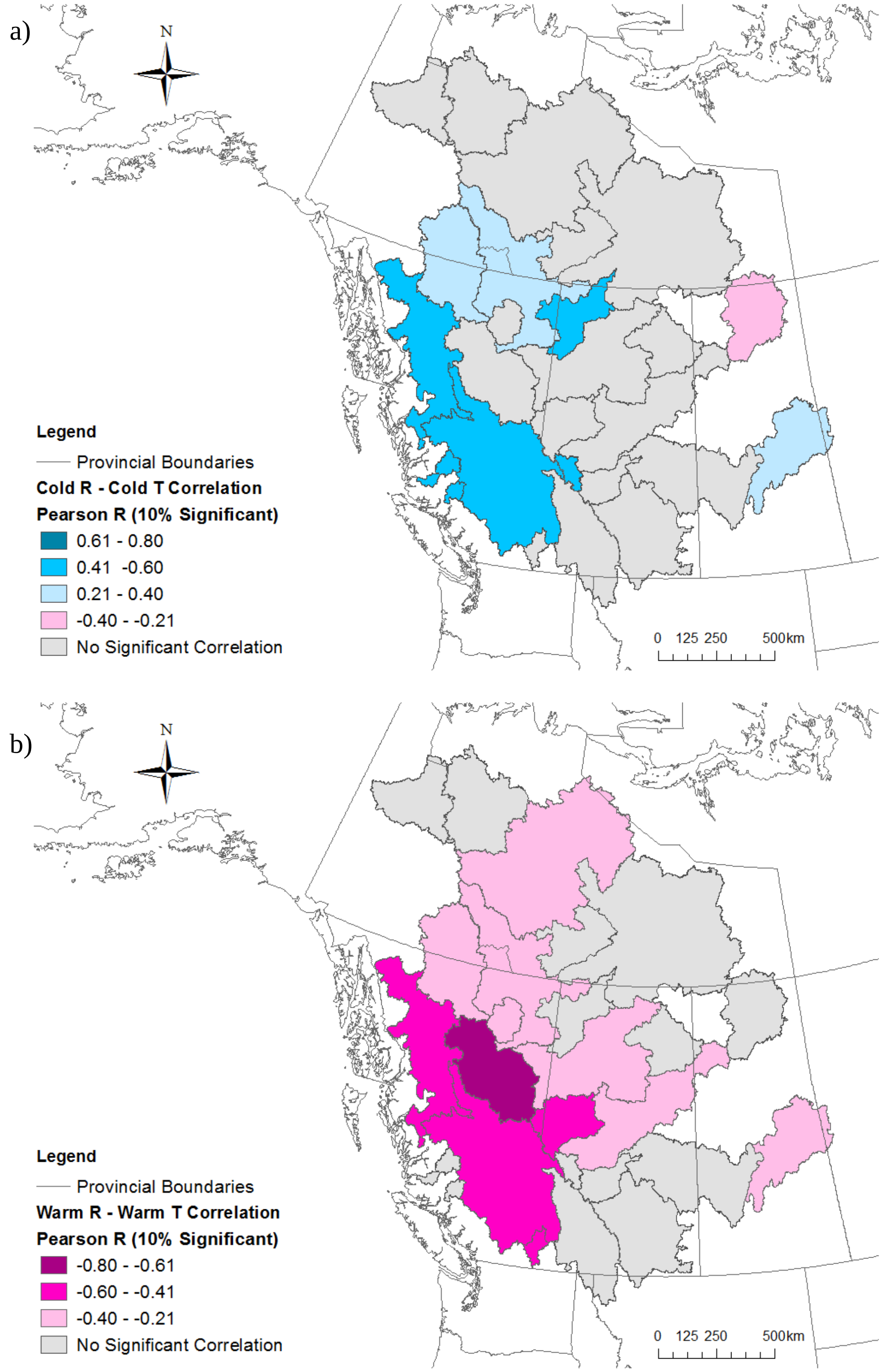
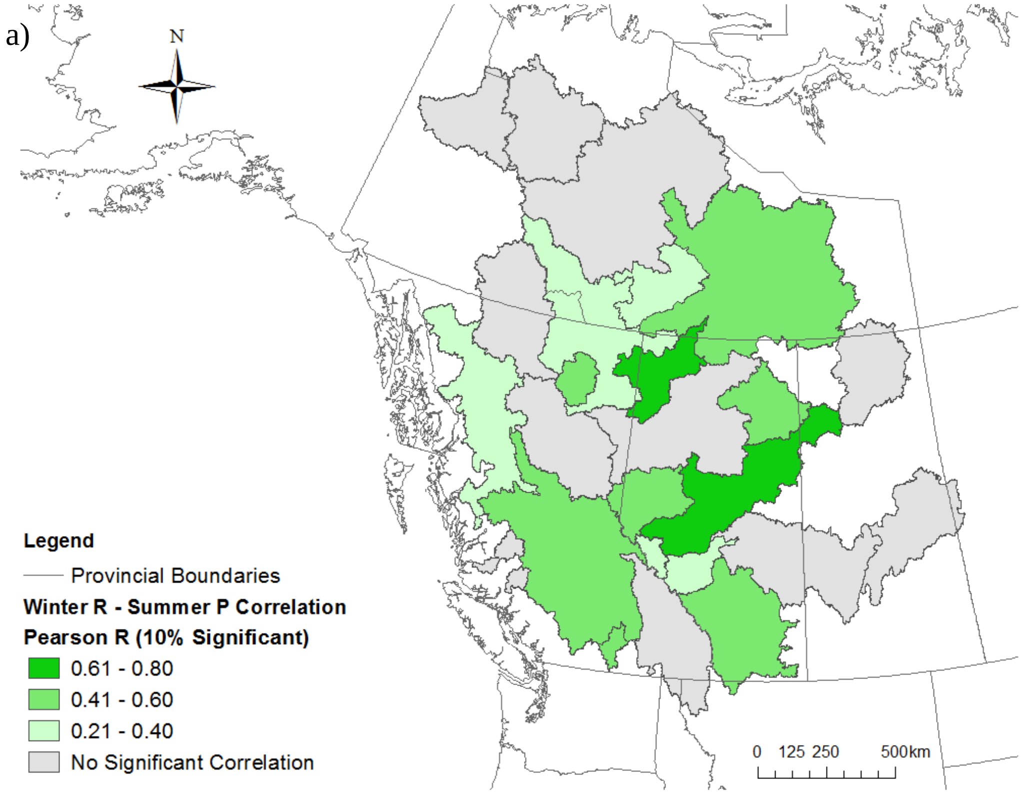


Figure D.3. Basins that exhibited significant correlations ($p < 0.1$) between a) cold season and b) warm season runoff and maximum temperature of the same seasons, respectively.



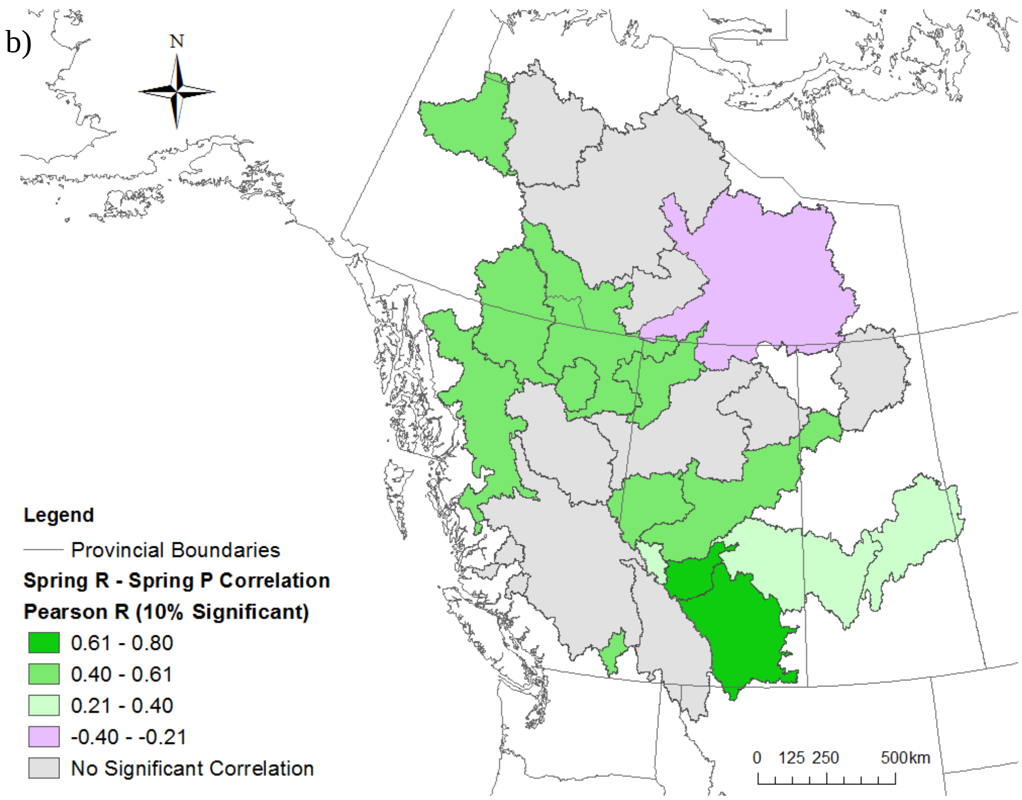


Figure D.4. Basins that exhibited significant correlations ($p < 0.1$) between a) winter runoff and summer precipitation and b) spring runoff and precipitation.

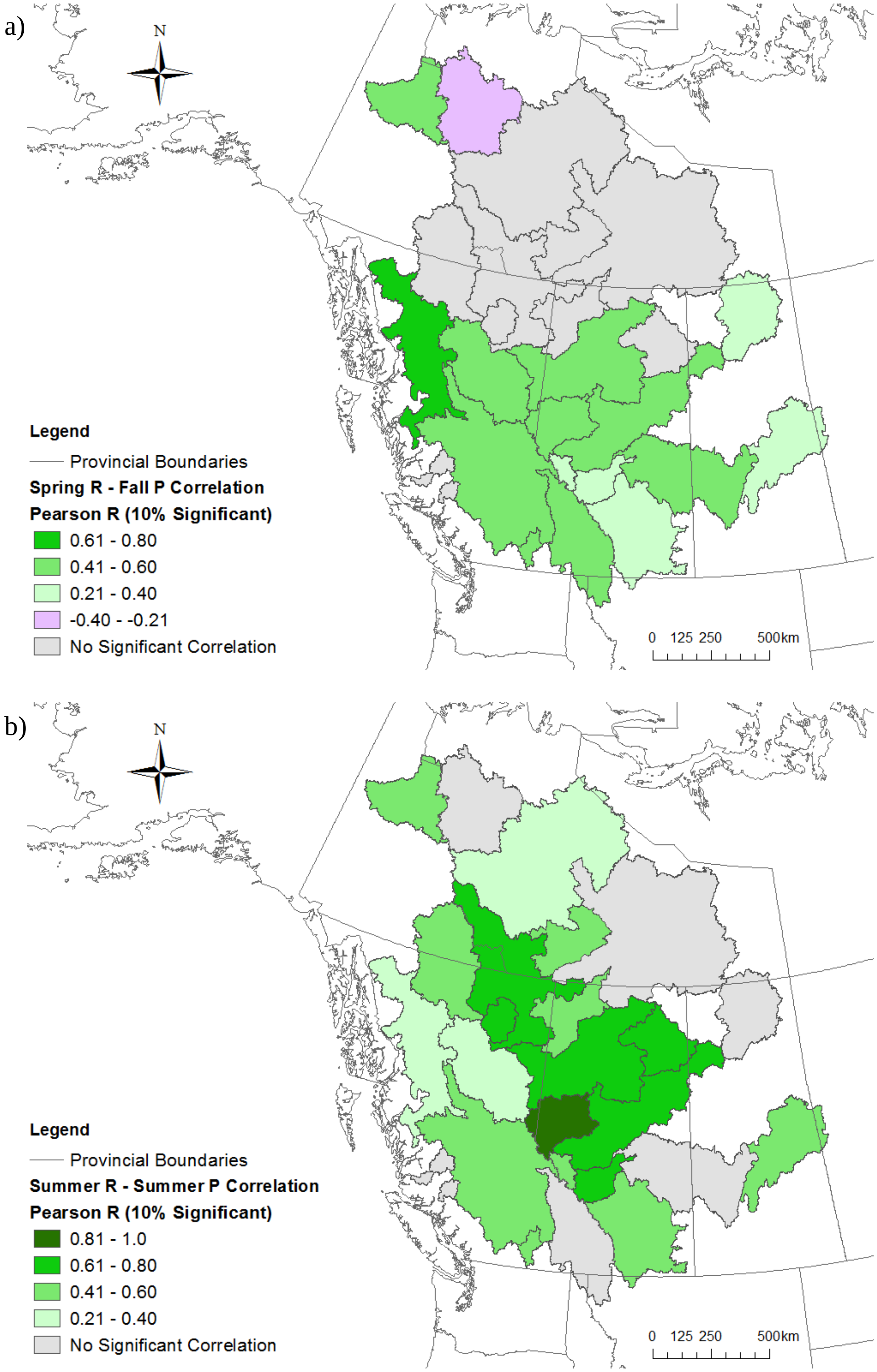


Figure D.5. Basins that exhibited significant correlations ($p < 0.1$) between a) spring runoff and fall precipitation and b) summer runoff and precipitation.

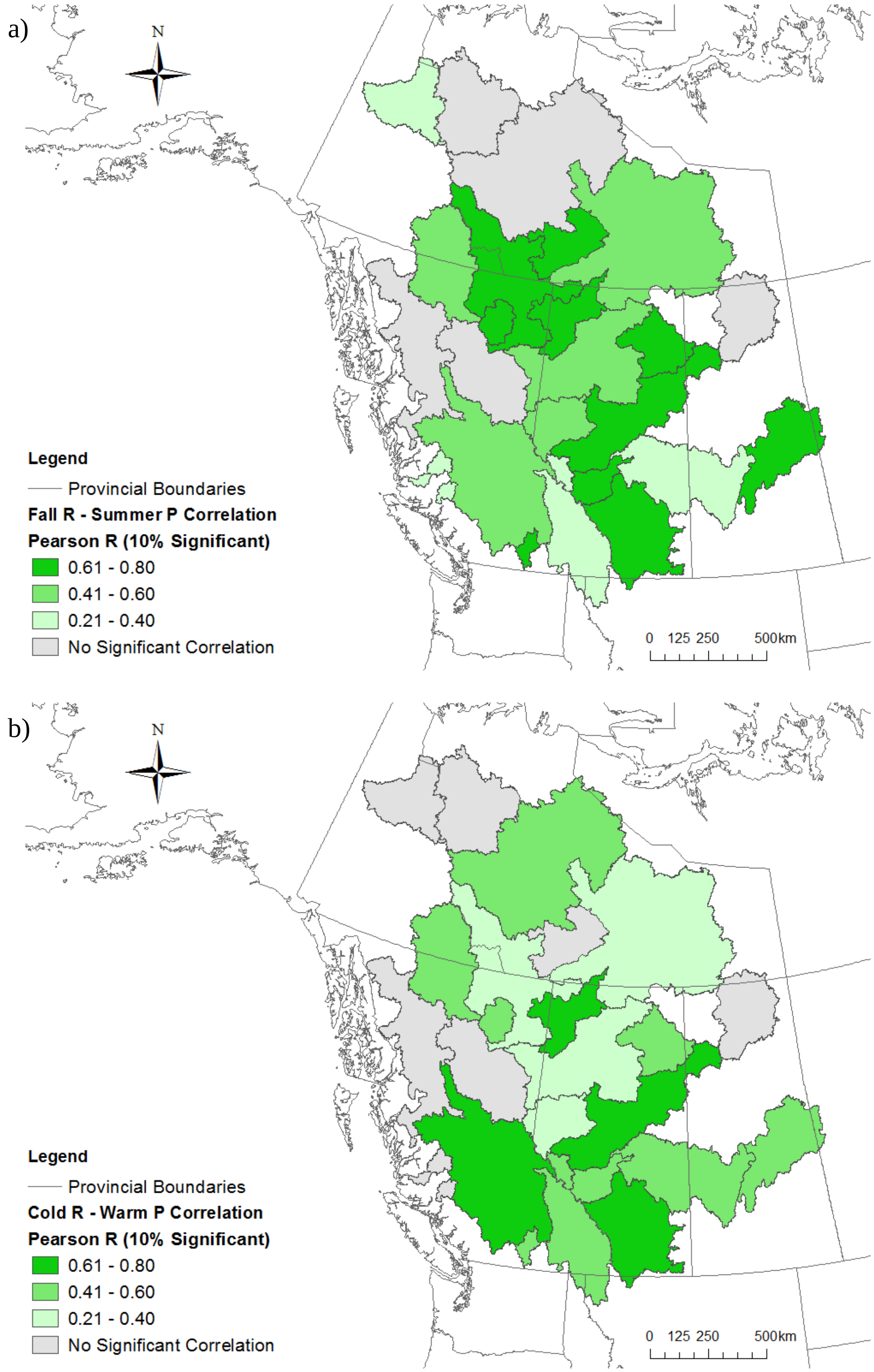


Figure D.6. Basins that exhibited significant correlations ($p < 0.1$) between a) fall runoff and summer precipitation and b) cold season runoff and warm season precipitation.

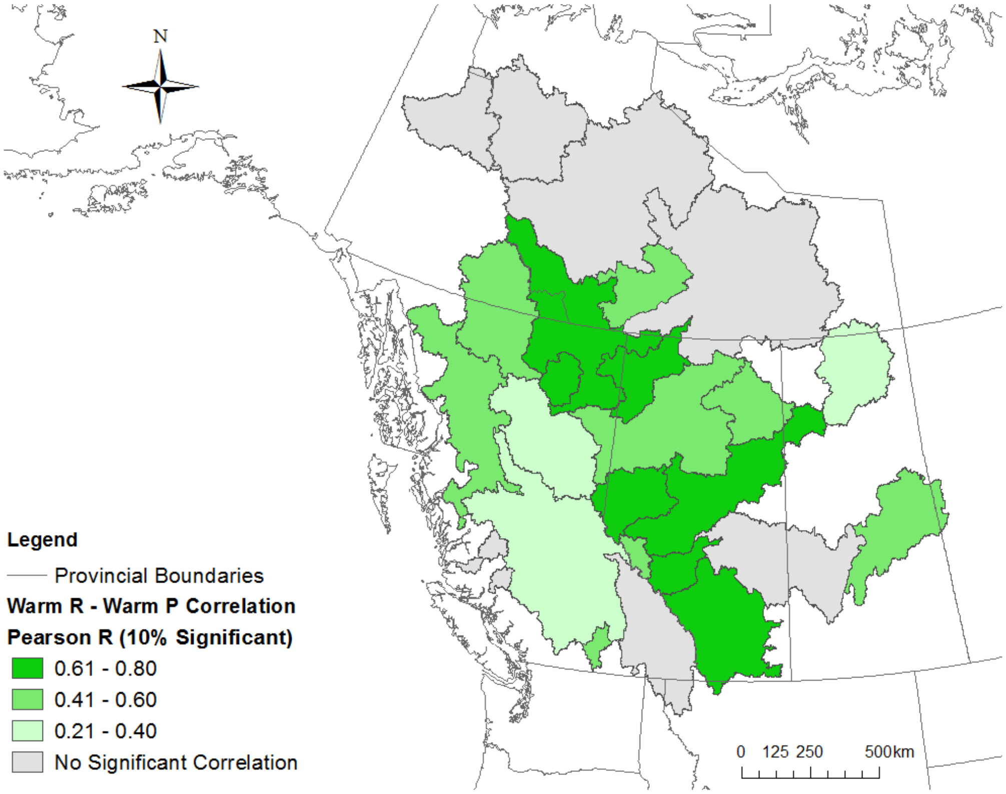


Figure D.7. Basins that exhibited significant correlations (p < 0.1) between warm season runoff and precipitation.