

Letter to the Editor – March 23, 2016

Dear Dr. Schaepli,

We again wish to express our thanks to the two anonymous reviewers for their time and effort on providing a subsequent review of the draft paper. Their comments have helped us in re-revising the paper and further strengthening it. Here we respond to the specific comments and explain the revisions we have made based on these comments.

Comments by Reviewer #1

“The authors have indeed carried out extensive updates to the literature review but their effort to integrate hydrology and hydrologic processes into the review comes almost as an afterthought in the final section of the paper which is very disappointing. It was also disappointing that the authors chose to defend the current structure of the paper by citing the number of downloads of the paper.

I was expecting at the very least some effort to highlight the hydrological links of the various climate variables in the beginning of each section. This would also have served to streamline the review and reduce the length of the paper. Take the air temperature section for example. The section starts with a discussion of the AHCCD dataset then goes on for 2 pages describing the trend methodology and results for numerous air temperature variables without any attempt to discuss the hydrological significance of the changes.

An effort to line-up key time-series as a multi-variate snapshot plot to focus the discussion might have been enough to edge the paper into the accept zone. As it stands, I do not see any compelling evidence that this paper offers an “integrated review” as claimed by the authors.”

Author’s response

We recognize, in part, the reviewer’s concern that our effort to integrate hydrology and hydrological processes into the review came almost as an afterthought. We have made several changes that should help to improve the integration throughout the paper (discussed below). However, we feel that a discussion of the hydrological links and significance of the changes is best placed at the end of the paper, after the changes in all variables have been reviewed and described. This is important since a key message of the paper is that the hydrological response to the observed changes has been complex and not easily inferred based on changes in any individual system component in isolation.

The reviewer noted that some effort to highlight the hydrological links of the different variables at the beginning of each section was expected. In response, we have added a short paragraph at the beginning of each section to provide some basic context and highlight the linkages in a conceptual sense. This is then built upon in the final section where we summarize the changes and how they are inter-related. Here we have made an important change to the structure of the paper. The previous version of the manuscript left this discussion to the very last sub-section, *concluding remarks*, which likely contributed to the appearance that hydrological linkages were only considered as an afterthought. We have now moved this discussion forward as a lead-in to the sub-section on process-level insights from the WECC observatories (following the recommendation of reviewer #2). This not only provides the discussion a better flow for the reader, but it also gives context for the more detailed, process-level studies and insights, and for the need to develop appropriate conceptual understanding. We have also added some further remarks to the final sub-section to underscore the significance in terms of understanding and projecting hydrological change in cold regions.

The reviewer suggests that a multi-variate comparison to focus discussion could have been included. We feel that this would add little that is not already clear from the previous figures and the discussion of sections 9.2 and 9.3, while it would increase the length of the paper. Rather, the focus of the discussion

has been improved from the structural change described above. In our revised paper, we no longer suggest this paper offers an “integrated” review, but instead a “comprehensive” review.

General Comments by Reviewer #2

“The revised manuscript is a substantial improvement over the original submission. Although the paper remains on the long side, I appreciate that the breadth of material precludes a shorter paper. I thank the authors for the additional synthesis (written as section 9.2), which I feel offers a logical bookend to the review material. Nevertheless, I offer one suggestion. Although the authors stress the point in the abstract that despite the generally robust changes in temperature, permafrost and snow cover/persistence, integrated streamflow change is variable (a function of complex processes and interplay). The paper would be that much more impactful if this point was stressed in the main text as well - perhaps as a lead-in to Section 9.2.

Author’s response

The reviewer has suggested that the paper would be more impactful by placing the discussion on systematic climatic and cryospheric change and complex hydrological response ahead of the discussion on insights from the observatories. We agree and have done this, as described above.

Specific Comments by Reviewer #2

Other technical points are as follows (page and line numbers refer to the tracked-changes version of the revised manuscript):

Page 5, Line 24: Is this a presumption, or an undisputed fact?

Response: We have changed the text to say “these alone do not explain the long-term changes” (p. 5, line 23 of the marked up re-submitted manuscript)

Page 6, Line 13: The use of “intensity” is ambiguous - are cold spells getting warmer or colder?

Response: We have clarified by removing the word “increase”, which was a mistake (p. 6, line 12)

Page 11, Line 24-27: This is a very awkward sentence - should be two sentences.

Response: We have made this change (p. 12, lines 13-14)

Page 17, Lines 13-15: Awkward sentence

Response: We have changed this to “The analysis was restricted to currently active stations, and excluded most higher order rivers as these are mostly affected by upstream regulation and other human influences (and are thus not part of the RHBN).” (p. 19, lines 7-9)

Page 19, Line 35-36: The meaning of “... observations at larger scales are directly reactivation of subsurface flowpaths.” is unclear.

Response: We have changed this to “In contrast, at the headwater and meso-scales (<1000 km²), these trends are rarely observed, creating some uncertainty as to whether activation of new and deeper subsurface flow pathways is the cause of change at larger scales.” (p. 22, lines 10-12)

Page 19, Line 38: What is it about thermokarsts that are changing? Increased frequency?

Response: We have added “thermokarst development” (p. 22, line 15)

We again thank you for your efforts as the handling editor and we look forward to hearing back from you.

1 **Recent climatic, cryospheric, and hydrological changes over the interior of western Canada: a**
2 **review and synthesis**

3
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9
10 ~~January~~March, 2016

11
12 **Abstract**

13
14 It is well-established that the Earth's climate system has warmed significantly over the past several
15 decades, and in association there have been widespread changes in various other Earth system
16 components. This has been especially prevalent in the cold regions of the northern mid to high-latitudes.
17 Examples of these changes can be found within the western and northern interior of Canada, a region
18 that exemplifies the scientific and societal issues faced in many other similar parts of the world, and where
19 impacts have global-scale consequences. This region has been the geographic focus of a large amount of
20 previous research on changing climatic, cryospheric, and hydrological ~~Earth system components~~regimes
21 in recent decades, while current initiatives such as the Changing Cold Regions Network (CCRN) introduced
22 in this review seek to further develop the understanding and diagnosis of this change and hence improve
23 the capacity to predict future change. This paper provides a ~~comprehensiven-integrated~~ review of the
24 observed changes in ~~variousthese~~ Earth system components and a concise and up-to-date regional picture
25 of some of the temporal trends over the interior of western Canada since the mid or late-20th century.
26 The focus is on air temperature, precipitation, seasonal snow cover, mountain glaciers, permafrost,
27 freshwater ice cover, and river discharge. Important long-term observational networks and datasets are
28 described, and qualitative linkages among the changing components are highlighted. Increases in air
29 temperature are the most notable changes within the domain, rising on average 2 °C throughout the
30 western interior since 1950. This increase in air temperature is associated with hydrologically important
31 changes to precipitation regimes and unambiguous declines in snow cover depth, persistence, and spatial
32 extent. Consequences of warming air temperatures have caused mountain glaciers to recede at all
33 latitudes, permafrost to thaw at its southern limit, and active-layers over permafrost to thicken. Despite
34 these changes, integrated effects on stream flow are complex and often offsetting. Following a review of
35 the current literature, we provide insight from a network of northern research catchments and other sites
36 detailing how climate change confounds hydrological responses at smaller scales, and we recommend
37 several priority research areas that will be a focus of continued work in CCRN. Given the complex
38 interactions and process responses to climate change, it is argued that further conceptual understanding
39 and quantitative diagnosis of the mechanisms of change over a range of scales is required before
40 projections of future change can be made with confidence.

41
42 **1. Introduction**

43
44 Recent warming of the Earth's climate system has been impacting many biogeophysical systems and their
45 interactions globally (IPCC, 2013). Changes have been particularly great in the northern high-latitudes,
46 where observations show shifts in the amount and phase of precipitation, diminishing seasonal snow
47 cover, retreat and loss of glaciers, warming and thawing of permafrost, earlier breakup of seasonal
48 freshwater ice cover, changes in the timing and magnitude of river discharge, and altered composition,

1 structure, and density of terrestrial vegetation communities (Serreze et al., 2000; ACIA, 2004; Hinzman et
2 al., 2005; White et al., 2007; Prowse et al., 2009b; Callaghan et al., 2011a; Derksen et al., 2012; AMAP,
3 2012; Bush et al., 2014). Responses to climatic and other environmental changes may be incremental or
4 alternately characterized by threshold-type behavior, often involving complex feedbacks, and there is
5 increased sensitivity to warming in areas with winter temperatures near freezing associated with the
6 phase change of water at 0 °C (e.g., Adam et al., 2009). Understanding past changes in these systems is
7 important, yet is difficult in part because of these complexities. Adding to the uncertainty, observational
8 datasets are generally limited to a relatively short period of record on the order of decades and there is
9 limited understanding of longer-term climate and environmental variability. Evaluating change across
10 datasets is challenging as data may not be homogeneous, it typically reflects different spatial and
11 temporal scales (e.g., in situ versus a satellite-derived average), and may be responding to different
12 processes depending on how and where measurements are collected. Anthropogenic factors such as land
13 and water management may also have a considerable impact (Nazemi and Wheeler, 2014) and confound
14 interpretation of Earth system change.

15
16 The interior of western Canada provides an immediate example of cold region environmental changes
17 observed globally and the societal issues faced in the context of such changes. Changes, including those
18 listed above, have been pervasive, while the costs associated with recent hydro-climatic extreme events
19 (e.g., floods, drought, and wildfire) have been increasing (e.g., Hanesiak et al., 2011; Pomeroy et al., 2015).
20 The principal continental-scale drainages, the Mackenzie and Saskatchewan River systems, support a
21 major area of Canada's food and energy production, mining, forestry, critical riverine and delta
22 ecosystems, growing cities, rural and aboriginal communities, and freshwater supply to the Arctic Ocean
23 and Hudson Bay. The region is highly vulnerable to climate change with pressures from natural resource
24 and hydroelectric development, irrigation demands, and population growth exacerbating the impacts
25 (Martz et al., 2007; MRBB, 2012). Consequently, climate and environmental change here are of concern,
26 not only at local and regional levels, but also at the global scale as this impacts the global natural resource
27 and food trades, and regional Earth system change influences the global climate system (RIFWP, 2013).

28
29 This area of Canada is the geographic focus of a major research initiative, the Changing Cold Regions
30 Network (CCRN; DeBeer et al., 2015; www.ccrnetwork.ca), which aims to improve the understanding,
31 diagnosis, and prediction of interactions among the cryospheric, ecological, hydrological, and climatic
32 components of the changing Earth system at multiple spatial scales over the Mackenzie and Saskatchewan
33 River basins (Fig. 1). The CCRN project was recently adopted as a Regional Hydro-Climate Project (RHP)
34 by the Global Water and Energy Exchanges (GEWEX) Hydro-Climate Panel. An early objective of CCRN is
35 to characterize observed Earth system changes across the interior of western Canada over the past several
36 decades, including an inventory and statistical analyses of change as observed from long-term federal and
37 provincial observational networks and other regional datasets. A network of local Water, Ecosystem,
38 Climate, and Cryosphere (WECC) observatories that span different environments within the domain
39 provides finer details and process-level insights into the observed changes (Fig. 1). Subsequent scientific
40 objectives of CCRN involve the development of improved diagnostic and predictive modelling tools, and
41 their application in better understanding this change and predicting interactions and feedbacks among
42 the changing Earth system components from local to regional scales.

43
44 There has been a substantial amount of previous work aimed at characterizing and quantifying recent
45 trends and variability in the climate and other Earth system components over this region. The CCRN builds
46 on a legacy of other preceding research initiatives, including the Mackenzie GEWEX Study (MAGS; Stewart
47 et al., 1998; Woo et al., 2008; www.usask.ca/geography/MAGS), the Boreal Ecosystem-Atmosphere Study
48 (BOREAS; Sellers et al., 1997; Hall, 1999), the Drought Research Initiative (DRI; Stewart et al., 2011;

1 www.drinetnetwork.ca), the Western Canadian Cryospheric Network (WC2N; <http://wc2n.unbc.ca>), the
2 International Polar Year (IPY), and the Improved Processes and Parameterization for Prediction in Cold
3 Regions Hydrology Network (IP3; www.usask.ca/ip3), among others. These major studies provided
4 important observations and insights into change, while also providing the foundation for further
5 investigations. Current parallel initiatives to the CCRN include the Canadian Network for Regional Climate
6 and Weather Processes (CNRCWP; www.cnrcwp.uqam.ca) and the Canadian Snow and Sea Ice Evolution
7 Network (CanSISE; www.cansise.ca). Considering this work, the aim of this paper is to bring together and
8 review the recent climatic, cryospheric, and hydrological changes over the interior of western Canada
9 documented in the literature, and to provide a concise and up-to-date regional picture of the recent
10 trends. Furthermore, we use insights from the WECC observatories to provide synthesis and guidance for
11 future research questions linking climate change to hydrological responses.

12
13 In the following sections we describe changes and trends in surface air temperature, precipitation,
14 seasonal snow cover, mountain glaciers and icefields, permafrost, freshwater ice cover, and river
15 discharge. The focus is generally on regional assessments of change based on extensive observing
16 networks, which provides context for more detailed local observations of change at CCRN WECC
17 observatories and elsewhere. Some principal observation networks and other important sources of
18 regional or long-term data for the detection of change are briefly described. We consider the issue of
19 distinguishing long-term trends from short-term variability or periodicity in records of limited length,
20 together with the role of large-scale, low-frequency oceanic–atmospheric oscillations in driving changes
21 over various time-scales. The paper concludes with some remarks on the quality and length of
22 observational datasets, highlights of some of the qualitative linkages among the observed changes in Earth
23 system components, and a short discussion on the complexities of climatic and cryospheric process
24 interactions and hydrological responses based on insights from local-scale observations and experimental
25 studies at some of the WECC observatories, ~~and highlights of some of the qualitative linkages among the~~
26 ~~observed changes in Earth system components~~. This provides the context for the diagnosis of change to
27 be pursued as subsequent work in CCRN, including the development of improved conceptual
28 understanding of process response and quantitative diagnostic modelling of these changes.

29 30 **2. Air Temperature**

31 32 2.1 Context

33
34 Surface air temperature and its diurnal and seasonal variability is a fundamental climatic element
35 characterizing a region, and affects nearly all other hydro-meteorological conditions, including
36 evaporation rate and precipitation phase and type. In cold regions, cryospheric and hydrological regimes
37 are strongly influenced by air temperature and its variation, especially around 0 °C in association with the
38 melting of snow and ice and the transition between snowfall and rain. Across the interior of western
39 Canada, mean annual air temperature (for the period 1981–2010) varies considerably, ranging from 5 to
40 10 °C in the southwest to –5 to –10 °C in the northern part of the region (Environment Canada, 2016). In
41 summer, average temperatures range from between 10 to 20 °C from north to south across the region,
42 while in winter these range from about –30 to –5 °C. During the spring period, the seasonal mean 0 °C
43 isotherm moves northward over a period of several months, beginning in March and early-April in the
44 south through until about late-May and June in the north, influencing the timing of snow and ice melt and
45 the spring freshet.

46 47 2.1.2 Adjusted and Homogenized Temperature Dataset for Canada

1 To facilitate research on climate and environmental change, the Climate Research Division of Environment
2 Canada has developed an Adjusted and Homogenized Canadian Climate Dataset (AHCCD;
3 www.ec.gc.ca/dccha-ahccd/). These data are based on federal monitoring stations across the country and
4 incorporate several adjustments to the original station records to address shifts from changes in
5 instrumentation and observing practices. Monthly adjustments were interpolated to each calendar day
6 to produce daily maximum and minimum temperatures, and in some cases observations from multiple
7 stations were joined to generate longer time series (Vincent et al., 2012). The current version contains
8 records for 338 Canadian locations, but station density and record length decrease considerably toward
9 the north, while the data availability over most of the Arctic and sub-Arctic parts of this region is restricted
10 to the mid-1940s to present (Fig. 2a). Based on these data, annual, seasonal, and monthly temperature
11 anomalies (departures from the 1961–1990 average) have been interpolated to a 50 km grid (CANGRD)
12 by Environment Canada, covering southern Canada from 1900 and the entire country from 1948
13 (<http://open.canada.ca/data/en/dataset/3d4b68a5-13bc-48bb-ad10-801128aa6604>). Rapić et al. (2015)
14 provide more details on the CANGRD product.

17 *2.2.3 Changes in Annual and Seasonal Air Temperatures*

19 Recent analyses based on the AHCCD indicate that mean annual air temperature trends at stations across
20 Canada have been dominated by statistically significant increases of about 1.5 °C between 1950 and 2010
21 (Zhang et al., 2000, 2011; Vincent et al., 2012, 2015). There is a strong spatial coherence in the trends,
22 with the strongest warming over western and northwestern Canada (1.5 to 3 °C). Night-time warming
23 (assessed from average daily minimum temperatures) has been slightly greater than daytime warming
24 (from average daily maximum temperatures) (Zhang et al., 2000; Vincent et al., 2012). The analysis of
25 Vincent et al. (2012) indicates that nationally the warmest year on record was 2010 (the last year of data
26 used in [thate](#) analysis), followed by 2006 and 1998.

28 To illustrate the spatial pattern and magnitude of recent trends in surface air temperature over western
29 Canada, we analyzed annual and seasonal CANGRD temperature data for the period 1950–2012. Trends
30 were derived following Zhang et al. (2000), with a first order autoregression process used to adjust the
31 temporal autocorrelations within the climate series. A two-step approach was used to estimate the
32 autocorrelation parameter (ϕ) and trend slope (β) iteratively and remove the autocorrelation from the
33 time series. Iterations were continued until the difference in the ϕ and β estimates in two consecutive
34 steps was less than 1%; the value of β was estimated from the de-autocorrelated time series based on the
35 method of Sen (1968). The P-value of the trend slope from the de-autocorrelated series was computed
36 using a rank trend test from Mann (1945) and Kendall (1975).

38 The results are shown in Fig. 3. Annual mean air temperature trends (Fig. 3a) show strong spatial
39 coherence with slightly greater warming in the northern areas (as found in other studies), with statistical
40 significance at the 95% confidence level at all grid points. On average, temperature over the region has
41 increased by just over 2 °C in the 63-year period, which exceeds the average increase over the global land
42 surface of 1.2 °C for the same period (based on the Global Historical Climate Network-Monthly (GHCN-M)
43 dataset available through the National Centers for Environmental Information;
44 www.ncdc.noaa.gov/climate-monitoring/). Seasonally, the greatest warming has occurred during winter
45 (Fig. 3c) and to a lesser extent, spring (Fig. ~~2b~~**3b**), while warming in the summer (Fig. 3d) and fall (Fig. 3e)
46 has been less pronounced with fewer statistically significant trends. In winter, the average temperature
47 increase over the region was 3.9 °C, with a maximum increase of up to 6 °C in parts of the northern
48 Mackenzie basin and surrounding areas. Comparison with the global-scale analysis of Hansen et al. (2010)

1 shows that winter warming here is among the highest of that worldwide. Consistent with this warming,
2 Bonsal and Prowse (2003) observed significant trends toward earlier spring 0 °C isotherm dates over the
3 region, ranging from 5 to as much as 20 days earlier over the latter half of the 20th century. From our
4 analysis, temperature increases during spring, summer, and fall over western Canada averaged 2.2, 1.2,
5 and 1.1 °C, respectively, all of which also exceed the corresponding global trends.

6
7 Large-scale modes of oceanic-atmospheric circulation influence surface air temperatures on various
8 timescales over western Canada, and thus factor in to the observed trends and interannual variability.
9 These include, for example, El Niño–Southern Oscillation (ENSO), the Pacific Decadal Oscillation (PDO),
10 the North Atlantic Oscillation (NAO) and Arctic Oscillation (AO), among others. Bonsal et al. (2001b) found
11 that ENSO and PDO influence winter temperature here, most strongly during El Niño episodes. At these
12 times, positive PDO phases were associated with strong positive temperature anomalies, while negative
13 PDO phases were associated with strong negative anomalies. They noted an increase in the occurrence
14 of El Niño events and predominantly positive PDO winters after 1976, which would, in part, account for
15 some of the observed winter warming. Vincent et al. (2015) quantified the component of seasonal and
16 annual temperature trends associated with the NAO and PDO over Canada during 1948–2012. Their
17 analysis confirmed that the PDO signal explained some of the observed trends in winter and spring,
18 accounting for between a few tenths of a degree up to 2 °C of the warming in some areas. However, after
19 removal of the influence of these indices, statistically significant trends were still observed. Bonsal and
20 Prowse (2003) found that despite a link between 0 °C isotherm dates and various indices such as ENSO,
21 PDO, NAO, and others, a relatively small amount of overall variance was explained. Although such large-
22 scale patterns influence regional temperatures, the period since the mid-20th century is sufficiently long
23 to capture the main (known) periodic effects, and these alone do not explain the long-term changes most
24 of the observed warming is due to other factors.

25 26 2.3-4 *Changes in Daily and Extreme Air Temperatures*

27
28 Not only have average annual and seasonal temperatures increased across the region, but major changes
29 in extremes such as maximum, minimum, and other percentiles of monthly and daily temperature have
30 been observed. Earlier work using the first generation AHCCD showed that over the latter half of the 20th
31 century most stations exhibited increasing trends in the lower and higher percentiles of daily minimum
32 and maximum temperature distributions, and there was a reduction in areas experiencing abnormal and
33 extreme cold conditions with a concomitant increase in areas experiencing abnormal and extreme warm
34 conditions (Zhang et al., 2000; Bonsal et al., 2001a). Bonsal et al. (2001a) noted this translates into fewer
35 days with extreme low temperature (mainly during winter, spring, and summer) and more days with
36 extreme high temperature (mostly winter and spring). They also reported a greater increase in the daily
37 minimum temperatures (i.e., greater nighttime warming), thereby reducing intra-seasonal standard
38 deviation of daily temperature. Vincent and Mekis (2006) examined a number of other indices, including
39 frost days, cold days, cold nights, warm summer days, warm days, warm nights, diurnal temperature
40 range, and standard deviation of minimum temperature (see their Table 1 for all definitions). They found
41 widespread reductions in the number of cold days and nights (maximum/minimum temperatures < 10th
42 percentile of the corresponding distributions) from 1950–2003 ranging from 10 to 50 days per year fewer
43 over the period, and to a lesser extent, increases in the number of warm days, nights, and warm summer
44 days. Other indices showed mostly mixed, non-significant trends. Mekis et al. (2015) examined trends in
45 extreme heat and extreme cold events (days with at least one hourly humidex value above 30 and with at
46 least one hourly wind chill value below –30, respectively) from 1953–2012 at 126 stations across Canada.
47 They found that extreme heat events had increased significantly at many of the stations across Canada
48 and extreme cold events had decreased significantly at virtually all stations. Wang et al. (2014) assessed

1 changes in one-in-20 year extreme temperatures (from annual maxima and minima of daily temperature
2 series) from 1961–2010, and found that warming was greatest for the extreme low temperatures and was
3 stronger in the north-western part of Canada. Little warming was observed in the extreme high
4 temperatures. They also reported that warming was stronger in winter than summer, and stronger during
5 nighttime than daytime, in accordance with Bonsal et al. (2001a).

6
7 Other studies have examined the variation in the intensity, duration, and frequency of cold and warm
8 spells (defined as 3 or more consecutive days with minimum/maximum temperatures below/above the
9 20th/80th percentile of the corresponding distributions). Shabbar and Bonsal (2003) found that winter
10 warm spells were increasing in frequency from 1950–1998 with up to 3 or 4 more events each year over
11 the period, and that these were becoming on average 1 to 4 days longer. Winter cold spells showed mixed
12 trends in frequency, but a clear decrease in duration (1 to 4 days shorter) and ~~increase in~~ intensity (3 to 6
13 °C warmer). Analysis of summer warm spells by Mekis and Vincent (2008) from 1950–2007 showed a
14 slight increase in frequency of about 2 events per year over the period. The influence of ENSO has been
15 found to have a role in the frequency of cold and warm spells. For example, Shabbar and Bonsal (2004)
16 reported El Niño events were associated with significant increases in the occurrence of winter warm spells
17 and the number of extreme warm days, and decreases in the occurrence of winter cold spells and the
18 number of extreme cold days, while the opposite was generally found for La Niña events.

20 3. Precipitation

22 3.1 Context

23
24 Like air temperature, precipitation is a fundamental climatic element characterizing a region that
25 influences the cryospheric and hydrological regime. For instance, the amount, timing, intensity, duration,
26 and phase of precipitation, along with other characteristics such as its seasonality and spatial and
27 elevational distribution control the formation and development of seasonal snowpacks (and in cases of
28 rain-on-snow, their melt), glacier accumulation, soil moisture and storage conditions in a watershed, and
29 ultimately hydrological responses in terms of runoff. Annual and seasonal mean precipitation totals vary
30 more strongly than temperature over western Canada, as influenced by latitude, elevation, distance from
31 moisture sources, and other factors. Total amounts vary from up to a few hundred millimeters per year
32 in the north and also parts of the interior plains of southwestern Canada, to many hundreds and up to
33 thousands of millimeters per year in the Rocky Mountains (Environment Canada, 2016). Snowfall
34 accounts for about 30% of total annual precipitation in the southern interior plains and up to more than
35 60% in the far north and in high elevation areas of the Rocky Mountains (Woo et al., 2008).

37 3.1.2 Adjusted Precipitation Dataset for Canada

38
39 As with surface air temperature, the Climate Research Division of Environment Canada has developed an
40 adjusted precipitation dataset for assessing changes and variability in Canadian precipitation
41 (www.ec.gc.ca/dccha-ahccd/). Careful adjustments were made for known measurement issues in the
42 station data for 464 locations across the country (Mekis and Vincent, 2011). Issues include wind
43 undercatch, evaporation and wetting losses, snow water equivalent (SWE) estimation from depth
44 measurement as influenced by variable snow densities, trace observations, and amounts accumulated
45 over several days. It is noted that measurement of solid precipitation in particular is highly problematic
46 and associated with large uncertainties in both raw data and corrected products. As with the temperature
47 data, there is a low density of stations in the North and the data availability is mostly limited to the period
48 since the mid-1940s here (Fig. 2b). Annual and monthly anomalies from the 1961–1990 baseline period

1 were expressed as normalized percentage departures and interpolated to the 50 km resolution CANGRD
2 by Environment Canada, covering southern Canada from 1900 and the entire country from 1948. Rapić
3 et al. (2015) found that CANGRD produced trends that were up to twice the magnitude of a multi-dataset
4 average in their evaluation of the consistency among various widely used gridded observation-based
5 climate datasets over the Canadian Arctic. This is related to the adjustment of the station precipitation
6 data for CANGRD, and shows the importance of including corrections in gridded climate datasets.

8 3.2.3 *Changes in Annual and Seasonal Precipitation*

9
10 In general, studies using the AHCCD have noted an increasing trend in the total annual precipitation over
11 most parts of western Canada since about 1950 (Zhang et al., 2000; 2011; Mekis and Vincent, 2011;
12 Vincent et al., 2015). To provide an up-to-date regional picture of the annual and seasonal trends, we
13 analyzed the CANGRD precipitation dataset over western Canada for the period 1950–2012 using the
14 same methodology as described above for air temperature. Figures 4 and 5, provide relative changes (as
15 a percentage of the average) and absolute changes, respectively. On average, annual precipitation has
16 increased by about 14% (50 mm) over the region since 1950 (Figs. 4a, 5a); however, there is considerable
17 variability in the magnitude and significance of local trends. Most of the increase has been in the North,
18 where precipitation has risen locally by as much as 60% (~200 mm). Caution needs to be used in
19 interpreting these trends, however, as some of the areas showing large increasing trends coincide with a
20 very low density of surface observing stations (Fig. 2b). In most other parts of the Mackenzie and
21 Saskatchewan River basins, the trends are not statistically significant and are low in magnitude with mixed
22 sign.

23
24 The seasonal precipitation trends also exhibit large variability across the region (Figs. 4b–e, 5b–e).
25 Broadly, the spatial patterns of trends in summer and fall, and to some extent spring, are similar. In
26 winter, there is a clear divide between increasing trends in the North and decreasing trends in the South.
27 In most of the northern Mackenzie basin, winter precipitation has increased by about 30 to 50%, while in
28 the southern Mackenzie basin and most of the Saskatchewan basin it has decreased by about 20 to 30%
29 (and as much as 50% or more in southern Alberta) (Fig. 4c). Absolute changes are mostly within about ±
30 30 mm, except over the southern mountain areas (Fig. 5c). Again, caution needs to be used as there is a
31 low density of stations in much of the North and most observing stations in the mountain areas tend to
32 be located at low elevation and may not be representative of higher elevation areas.

33
34 In addition to changes in the amount of precipitation, there have also been observed shifts in its phase.
35 Zhang et al. (2000) and Vincent et al. (2015) examined trends in the ratio of annual and seasonal snowfall
36 to precipitation totals; from 1948–2012, this ratio decreased over southwestern Canada (from 0–15%)
37 and increased over much of northern Canada (by 5–20%). The greatest changes occurred in spring over
38 the western half of the country, with widespread reductions of up to 20% or more, reflecting the effects
39 of warmer temperatures. Mekis and Vincent (2011) separately examined annual and seasonal trends in
40 both rainfall and snowfall. They found that over the past 60 years, rainfall totals have increased annually
41 and in all seasons with the most pronounced change during spring, with increases of between 30 and 50%
42 over much of western Canada. Annual snowfall amounts had decreased across most of southwestern
43 Canada (reductions of 10 to 30%) but increased in much of the North (by up to 30%), with the greatest
44 changes occurring during winter and spring (reductions of up to 50% in the southwest and increases up
45 to 50% for some northern stations).

46
47 Precipitation variability here is linked to large-scale modes of atmospheric-oceanic circulation, including
48 ENSO, PDO, NAO, and others. El Niño and La Niña events tend to be associated with distinct negative and

1 positive winter precipitation anomalies over southwestern Canada (Shabbar et al., 1997; Gan et al., 2007),
2 while positive/negative phases of the PDO tend to be associated with drier/wetter than normal winters
3 (Whitfield et al., 2010). Bonsal and Shabbar (2008) and Whitfield et al. (2010) provide useful summaries
4 of the influence of these modes on the hydro-climatology of western Canada. While they play an
5 important role in the inter-annual to inter-decadal variation of precipitation, longer-term trends appear
6 to be mainly independent of this influence. For example, Vincent et al. (2015) assessed trends in annual
7 and seasonal total precipitation and snowfall ratio after removal of the influence of PDO and NAO indices
8 and found that the combination of these indices explained less than 10% of the trends over 1948–2010
9 and 1900–2012.

10 11 *3.3-4 Changes in Daily and Extreme Precipitation*

12
13 Aspects of the seasonality, frequency, duration, and intensity of precipitation events have changed over
14 the region in recent decades. For the period 1950–2003, Vincent and Mekis (2006) reported more days
15 with precipitation and especially with rainfall (10 to 30 days per year over the period), and a decrease in
16 average daily intensity (i.e., total annual precipitation divided by number of days with precipitation; 0 to
17 3 mm d⁻¹). Other indices characterizing extreme heavy precipitation events and maximum annual dry
18 spells did not show clear patterns or statistically significant trends. Stone et al. (2000) used station-
19 dependent thresholds to classify event intensity and found that during the latter half of the 20th century,
20 the frequency of lighter events decreased while that of intermediate and heavier events increased,
21 respectively by up to a few percent per decade. Zhang et al. (2001b) defined heavy rainfall and snowfall
22 events for each season using a threshold value that is exceeded by an average of 3 events per year.
23 Temporal variations of regional heavy precipitation displayed strong inter-decadal variability with limited
24 evidence of long-term trends over the latter part of the 20th century, except in the number of heavy
25 snowfall events in fall and winter, which increased over all of northern Canada. Mekis et al. (2015) found
26 little evidence of changes in heavy rainfall events (accumulated rainfall >10 mm, 25 mm, and 50 mm over
27 periods of 1 hour, 24 hours, and 48 hours, respectively) for the region over the period 1960–2012. They
28 noted that there is no apparent regional pattern in such extreme events because they are highly localized
29 and the station density is relatively low. Looking at the hydrological character of rainfall, Shook and
30 Pomeroy (2012) examined trends in short duration convective events, multiple day accumulations, and
31 rainfall occurring during the spring and fall over the Canadian prairies. Over the periods 1901–2000 and
32 1951–2000, the fraction of summer rainfall from convective events has decreased at many locations, while
33 that from multiple-day events has increased significantly.

34 35 *3.4-5 Drought*

36
37 There have been a number of severe prairie droughts documented over the instrumental record, with
38 multi-year droughts occurring in the 1890s, 1930s, late 1950s and early 1960s, 1980s, and 2000s (Bonsal
39 and Regier, 2007, Bonsal et al., 2011). In the first half of 2015, much of western Canada was experiencing
40 abnormal to record dry conditions—in many areas immediately following a several year period of
41 historical record wetness and flooding. This led to widespread forest fires, low water levels in streams,
42 lakes and reservoirs, and low soil moisture levels, and was unusual in that it stretched over a vast area
43 from Mexico to Alaska. Since the beginning of the 20th century there has been decadal-scale variability in
44 drought occurrence as indicated by various precipitation and soil moisture indices, but there has been no
45 consistent long-term trend in drought frequency or magnitude (Millett et al., 2009; Qian et al., 2010;
46 Bonsal et al., 2013). This variability has tended to coincide primarily with precipitation variations
47 modulated by large-scale modes of oceanic-atmospheric circulation (Bonsal et al., 2011; Shabbar and
48 Skinner, 2004; Bonsal and Shabbar, 2008). Through the use of proxy information it appears that extended

1 drought conditions during the 20th century have been relatively mild in comparison to the pre-settlement
2 era on the prairies, and there is evidence of climate-driven non-stationarities in hydrological variables
3 over the past several centuries or millennia (Bonsal et al., 2013; Razavi et al., 2015). Bonsal et al. (2011)
4 provides a useful review of drought research in Canada. The recently completed Drought Research
5 Initiative (DRI) network was established to conduct a comprehensive study of the severe 1999–2005
6 Canadian prairie drought (Hanesiak et al., 2011; Stewart et al., 2011).

8 **4. Seasonal Snow Cover**

10 4.1 Context

12 The seasonal snow cover that forms over the landscape during the winter months is an important feature
13 influencing the climate, hydrology, and ecology of cold regions and mountainous areas (Male, 1980; Gray
14 and Male, 1981). For instance, it has a strong control on the surface energy exchange with the
15 atmosphere, represents an important storage of water that is released during the spring melt period, and
16 affects ground thermal regimes and surface vegetation through its insulating properties. Changes and
17 variability in snow cover therefore have a large impact on many other Earth system processes in these
18 regions, and feedbacks with the climate system can intensify local and regional patterns of change and
19 have global implications (Callaghan, et al., 2011b). Snowpack evolution is closely linked with climatic
20 elements such as air temperature and precipitation, and interannual variations and trends in snow cover
21 conditions coincide strongly with those of such climate variables.

23 4.1-2 Snow Cover Datasets

25 The Meteorological Service of Canada (MSC) has produced the Canadian daily snow depth database and
26 the Snow Water Equivalent (SWE) Database, based on in situ observations of surface snow cover at
27 climatological stations across Canada and at snow course locations (MSC, 2000). Brown and Braaten
28 (1998) describe the snow depth database, including quality control procedures for internal consistency,
29 the effects of station shifts and urban warming, and the reconstruction of missing values. The data
30 represent about 400 stations with varying record lengths, few of which began before the mid-1940s, while
31 the spatial density and record length tend to decrease considerably in the North. The measurements also
32 tend to be biased to low elevations and open areas (Brown and Braaten, 1998). The SWE database
33 contains weekly, biweekly, or monthly measurements by a number of agencies, but since the late-1990s
34 it has not been actively maintained; in many cases, updates to snow course data can readily be obtained
35 from the various provincial/territorial agencies involved (MSC, 2000). The data primarily cover the period
36 from about 1950 to the mid-1990s, with a pronounced decline after 1985.

38 Remotely sensed snow cover datasets provide a useful source of information, supplementing in situ
39 observations and extending coverage over broad regions (Hall et al., 2006). Brown et al. (2010), Frei et al.
40 (2012), and Kelly (2012) describe some of the more widely used satellite and model-derived snow cover
41 products. The longest time series is the National Oceanic and Atmospheric Administration (NOAA) weekly
42 snow cover product, which includes near-consistent snow cover mapping since 1966 (Robinson et al.,
43 1993). NASA's Moderate Resolution Imaging Spectroradiometer (MODIS) also provides a range of
44 valuable snow cover products, although data availability is only from 1999 (Hall et al., 2002; Hall and Riggs,
45 2007). Coarse resolution and classification thresholds lead to uncertainty in these products, particularly
46 in mountain regions (e.g., Brasnett, 1999; Brown et al., 2010). Passive microwave sensors such as the
47 Scanning Multichannel Microwave Radiometer (SMMR), Special Sensor Microwave/Imager (SSM/I), and
48 Advanced Microwave Scanning Radiometer, EOS (AMSR-E) provide information on snow depth, SWE, and

1 melt onset, but are limited by variations in snowpack physical properties and wetness that affect the
2 signal, low spatial resolution that does not capture local-scale accumulation, and restrictions associated
3 with both shallow or intermittent snow, and deep snow (>120–150 mm SWE) (Frei et al., 2012; Kelly,
4 2012).

6 4.2.3 *Changes in Snow Cover*

8 Over most of Canada there has been a pattern of decreasing snow depths and snow cover duration and
9 extent since the mid-1970s, with the largest declines in western Canada and proportionally greater
10 changes later in winter and spring (Brown and Braaten, 1998; Dyer and Mote, 2006; Derksen et al., 2008;
11 Derksen and Brown, 2012). Brown and Braaten (1998) analyzed the Canadian daily snow depth database
12 for the period 1946–1995 and found widespread and spatially coherent decreases in depth that increased
13 in magnitude and spatial extent from January through March. Maximum changes were found over
14 western Canada, where reductions of between 1.0 to 1.5 cm year⁻¹ were observed. Decreases during the
15 fall were not as widespread or as great in magnitude. Analyses of remotely sensed data support these
16 observations and show that interannual variations of snow cover extent are highly correlated over broad
17 regions in western Canada (Frei and Robinson, 1999; Robinson and Frei, 2000). Using the NOAA weekly
18 snow cover product, Déry and Brown (2007) reported that over the period 1972–2006, the weekly mean
19 trend in snow cover extent was $-0.78 \times 10^6 \text{ km}^2 \text{ (35 years)}^{-1}$ for North America, and that the trends were
20 amplified further northward, consistent with a surface albedo–snow cover extent feedback reinforcing
21 the anomalies. Brown et al. (2010) analyzed 10 separate snow cover data sources covering different
22 periods between 1967 and 2008, and showed that over this period, May and June snow cover extent have
23 exhibited a considerable decline, decreasing by 14% and 46% respectively across the pan-Arctic region,
24 primarily as a result of earlier snowmelt and snow cover depletion. Several studies have shown trends
25 toward both earlier melt onset and shorter overall melt duration over North America since at least 1960
26 (Dyer and Mote, 2007; Tedesco et al., 2009). Derksen and Brown (2012) used the NOAA snow cover record
27 and showed that successive records for the lowest North American June snow cover extent were set in 3
28 of the last 5 years between 1967 and 2012.

30 To illustrate the pattern over western Canada, Fig. 6 shows recent trends in snow cover duration from the
31 Canadian daily snow depth dataset (1950–2012; Fig. 6a) and the NOAA weekly snow cover product (1972–
32 2013; Fig. 6b). Trends were computed using the method of Zhang et al. (2000) described above, with
33 significance assessed at the 95% confidence level. The magnitude of trends in annual snow cover duration
34 have ranged from about 1–15 fewer days per decade, with an average decline of about 4 days per decade,
35 almost entirely due to reductions occurring during the spring season (Brown et al., 2010; personal
36 communication, R. Brown, 2014). This equates to a shortening of the snow cover period from 1–2 months
37 over the region since 1950. The results in Fig. 6a and b differ somewhat in magnitude and spatial pattern,
38 due in part to differences in datasets and temporal periods, but generally point to greater changes in the
39 southern and western parts of the Mackenzie and Saskatchewan River basin, and lesser change in the
40 northeastern Mackenzie basin (also reported by Derksen et al. (2008)). It is noted that snow cover trends
41 in the mountain regions from both in situ and remotely sensed data are generally more uncertain due to
42 the low elevation bias of climate stations and the coarse resolution of satellite imagery.

44 At regional to continental scales, snow cover trends are mostly associated with pervasive climate warming
45 patterns independent of low-frequency atmospheric circulation patterns and teleconnections (Derksen
46 and Brown, 2012). Ge and Gong (2009) noted that at these scales, the regional domain of climate mode
47 teleconnections is exceeded. More locally, certain modes of variability (e.g., ENSO, PDO, NAO) influence
48 the interannual variability of snow cover (Derksen et al., 2008; Ge and Gong, 2009; Bao et al., 2011), but

1 are not the main driving factor behind the decreasing trends in western Canada. For instance, Vincent et
2 al. (2015) noted that trends in various snow cover indices from the Canadian daily snow depth dataset
3 were almost identical after removal of the influence of PDO and NAO.

4
5 Detecting regional changes in SWE is more challenging due to difficulties in estimating variable snow
6 density and problems retrieving SWE information from microwave sensors, and robust information on
7 large-scale, long-term SWE trends from satellite data is not available (Kelly, 2012). Brown (2000) used a
8 combination of in situ snow course and synoptic station snow depths, along with NOAA weekly snow
9 cover data to reconstruct monthly SWE for the period 1915–1997 over North America. He found overall
10 statistically significant increasing trends in SWE for December, January, and February, and a significant
11 decrease in April, but did not provide details on regional patterns. Some work has examined SWE
12 variations over western Canada using SMMR and SSM/I data (Walker and Silis, 2002; Derksen et al., 2002,
13 2003; Tong et al., 2010), but the analysis periods were relatively short and clear trends could not be
14 detected. Derksen et al. (2004) combined SMMR and SSM/I with in situ SWE observations over the
15 provinces of Saskatchewan and Manitoba and several states in north-central USA to examine SWE
16 variability from 1915–2002. Derksen et al. (2008) used in situ observations, NOAA weekly snow cover,
17 and SMMR and SSM/I SWE retrievals to examine trends over the Mackenzie River basin from 1950–1999.
18 In both studies, monthly SWE in spring did not show clear long-term trends, but rather indicated decadal-
19 scale variability that varied regionally. Squires (2014) focused on a more limited part of the Mackenzie
20 basin around the Great Slave Lake region using snow course data collected by Aboriginal Affairs and
21 Northern Development Canada to assess trends in snow depth and SWE from the mid-1960s up to 2010.
22 They reported progressively earlier onset of snowmelt across most of this area, along with regionally
23 varying changes in maximum SWE.

24 25 **5. Mountain Glaciers and Icefields**

26 27 5.1 Context

28
29 Mountain glaciers represent a major geomorphic agent shaping the landscape, they directly affect local
30 surrounding climatic conditions, and they provide a source of runoff well beyond the spring snowmelt
31 freshet period (Moore et al., 2009). Glacier mass balance responds directly to variations in snow
32 accumulation and snow and ice ablation, as influenced by air temperature, precipitation, radiation, and
33 other meteorological variables, while extended perturbations in the mass balance tend to produce a
34 delayed response in glacier geometry, surface elevation, and terminus position (Oerlemans, 1989). Thus,
35 patterns of glacier behavior over time provide a proxy indicator of the local to regional prevailing climatic
36 conditions. Globally, mountain glaciers have predominantly been losing mass and retreating, and there
37 is concern that as this continues their ability to provide a reliable source of meltwater runoff during the
38 summer and fall period will be reduced (Radić and Hock, 2014). In western Canada, glacier runoff has
39 been shown to provide a buffering mechanism during extreme dry years (Hopkinson and Young, 1998;
40 Comeau et al., 2009), but as glacier cover diminishes, they may no longer be able to effectively augment
41 low flows (Demuth and Pietroniro, 2003; Marshall et al., 2011).

42 43 5.1.2 Glacier Inventories and Mass Balance Data

44
45 Glaciers and icefields occur extensively throughout the mountain regions of western Canada, covering an
46 area of roughly 50,000 km² and including almost 12,000 individually documented ice masses (Ommanney,
47 1980; 2002a; Fig. 7). The Canadian Glacier Inventory project was initiated as an International Hydrological
48 Decade (IHD) contribution in the late 1960s and led to the production of the Glacier Atlas of Canada, which

1 includes detailed maps identifying all known glaciers, catalogued by Water Survey of Canada drainage
2 basins. Ommanney (1980; 2002b) describe the inventory and the procedures used; digital versions of the
3 original maps can be accessed through the GeoGratis website (<http://geogratis.cgdi.gc.ca/>). More recent
4 work has updated and extended this inventory over parts of western Canada, contributing to the Global
5 Land Ice Measurements from Space (GLIMS) project (Bolch et al., 2010; Demuth et al., 2014).

6
7 Detailed in situ observations and mass balance records are available only for a very limited number of
8 glaciers over the region. Ommanney (2002a) and Østrem (2006) provide accounts of the history of
9 glaciological investigations in western Canada and the selection of reference mass balance glaciers. The
10 most well-documented and studied glacier in the region is Peyto Glacier (Fig. 7), which has been
11 documented since the late 1800s and has near continuous annual and seasonal mass balance records
12 going back to 1965, beginning as part of the IHD activities (Østrem, 2006; Demuth and Keller, 2006). In
13 contrast, monitoring activities at most other glaciers have been short-lived and are largely discontinued,
14 while a concern for ongoing initiatives is the retreat and disintegration of glacier termini, making access
15 difficult and restricting continuity of measurement programmes.

17 5.2.3 Glacier Changes

18
19 Glaciers across western North America have predominantly retreated over the past century and the
20 observational records have shown mostly negative net mass balances (Moore et al., 2009). As shown in
21 Fig. 8, the net annual mass balance of Peyto Glacier has exhibited strong inter-annual variations but has
22 been mostly negative since 1965 (average -580 mm water equivalent (WE)); from 1965–2012 the
23 cumulative mass balance has indicated a loss of over 27 m WE averaged over the glacier (Demuth and
24 Keller, 2006; personal communication, M. Demuth, 2014). Demuth and Keller (2006) noted that the
25 annual mass balance is mainly driven by variations in winter balance, and that a shift to primarily negative
26 net balance after 1976 was driven mainly by a reduction in winter snowfall in association with the shift in
27 the PDO index to warm phase. Since the end of the Little Ice Age (LIA) in the 19th century—when most
28 glaciers were at their maximum Holocene extent—Peyto Glacier has retreated by about 3 km, and in
29 recent years a large pro-glacial lake has formed at its terminus. Most valley glaciers in the Rocky
30 Mountains have retreated by 1 to 2 km since the end of the LIA (Ommanney, 2002c).

31
32 The rates of retreat have varied, with initially rapid rates in the first half of the 20th century, followed by a
33 short-lived period of stabilization or advancement until 1980, and continuing retreat since then (Moore
34 et al., 2009). The records at the Athabasca and Saskatchewan Glaciers, together with observations
35 elsewhere in the Rockies, show a sharp decline in glacier recession in the 1960s and 1970s, attributed to
36 a short-lived period of positive net mass balance (Luckman, 1998). McCarthy and Smith (1994)
37 documented historical glacier length variations within the southern Rockies and showed that individual
38 glaciers displayed a wide range of patterns, with some stabilizing after the 1950s, others slowing in their
39 rate of retreat, and still others exhibiting an accelerated recession beginning in the last few decades of
40 the 20th century.

41
42 Glacier area changes have been investigated in many recent studies, and at the regional scale have been
43 reported to have declined from as little as 11% over 53 years to as much as 29% over 13 years, with most
44 studies reporting about a 25% decline in glacier cover since the mid- to late-20th century (Hopkinson and
45 Young, 1998; DeBeer and Sharp, 2007; Schiefer et al. 2007; Demuth et al., 2008; 2014; Jiskoot et al., 2009;
46 Bolch et al., 2010; Tennant et al., 2012; Tennant and Menounos, 2013; Beedle et al., 2015). Individually,
47 glaciers have exhibited a wide range of local changes from small net advances to complete disappearance,
48 reflecting the strong control of local factors on glacier mass balance, dynamics, and response. A number

1 of features have been commonly reported in western Canada, including: 1) glaciers have generally been
2 continuing to lose mass and retreat, and rates of loss have accelerated in the last few decades; 2) smaller
3 glaciers have tended to exhibit greater variability in their relative area changes, while larger glaciers have
4 exhibited more consistent and moderate relative changes; and 3) collectively, most of the glacier area loss
5 has been due to the retreat of larger glaciers that comprise a greater proportion of the regional ice cover.

6
7 Volumetric changes have been measured through comparison of repeat digital elevation models (DEMs),
8 which has indicated thinning rates from about 0.5 m to 0.9 m per year (Schiefer et al., 2007; Tennant and
9 Menounos, 2013). Hopkinson and Demuth (2006) used repeated airborne Lidar (Light detection and
10 ranging) measurements at Peyto Glacier and found that in a period of just under 2 years, the glacier's
11 surface had thinned by about 3.5 m on average and the glacier had lost $33 \times 10^6 \text{ m}^3$ of ice. Schiefer et al.
12 (2007) found that glacier volume in the Rockies declined by about 17 km^3 over the period 1985–1999.
13 Marshall et al. (2011) used an empirical volume–area scaling relationship (e.g., Chen and Ohmura, 1990)
14 and volume approximations based on glacier surface slope, and estimated the present volume of glaciers
15 in the eastern slopes of the Rocky Mountains to be $55 \pm 15 \text{ km}^3$, corresponding to an average glacier
16 thickness of 57 m.

17 18 5.3.4 Contributions to River Discharge

19
20 In western Canada, several recent studies have compared glacier losses with observed discharge and used
21 various modelling approaches to estimate the proportional contribution by glaciers (Hopkinson and
22 Young, 1998; Comeau et al., 2009; Marshall et al., 2011; Jost et al., 2012; Naz et al., 2014; Bash and
23 Marshall, 2014). Glacier wastage, referring to ice loss as a result of any negative net mass balance, has
24 been found to account on average for about 1–5% of the annual discharge of the larger rivers exiting the
25 eastern slopes of the Rockies, and up to 10% or slightly more of the summer (July–September) flow
26 (Comeau et al., 2009; Marshall et al., 2011; Bash and Marshall, 2014). In an extreme low flow year,
27 Hopkinson and Young (1998) found that glacier wastage supplied 13% of the annual flow of the Bow River
28 at Banff, with a maximum monthly contribution in August of 56%. Jost et al. (2012) found that glacier ice
29 melt (excluding seasonal snow cover) supplied an average of 6% of the annual flow of the upper Columbia
30 River between 1970 and 2007, which increased to as much as 25% and 35% of August and September
31 flow. Naz et al. (2014) simulated glacier contributions to the upper Bow River basin over 1981–2007. Their
32 results indicated that while summer and annual glacier melt showed increasing tendencies, summer and
33 annual discharge showed decreasing tendencies. They speculated that the decline in discharge is
34 associated with a combined effect of decreases in glacier cover and precipitation. Demuth and Pietroniro
35 (2003) observed that the discharge from glacierized catchments in the North Saskatchewan River basin
36 during the transition-to-baseflow (TBF) period has decreased since the mid-20th century, despite
37 increasing precipitation during the TBF period and greater ice melt. This response is commensurate with
38 reductions in glacier area. They also observed an increase in the variability of discharge during the TBF
39 period, pointing to a reduction in the buffering capacity of glaciers in this system.

40 41 **6. Permafrost**

42 43 6.1 Context

44
45 Permafrost, defined as ground that remains at or below a temperature of 0 °C for at least two consecutive
46 years, is a pervasive feature across approximately half of the Canadian land mass (Smith, 2011). At its
47 southern limits in Canada it occurs as sporadic and discontinuous permafrost with annual temperatures
48 near 0 °C, closely corresponding with the mean annual air temperature isotherm of 0 °C (Burn, 2012), and

1 it transitions to continuous, deep, and colder permafrost with increasing latitude (Fig. 9). The seasonal
2 thawing of the upper active layer is linked with air temperature during the summer as well as the
3 preceding winter (Smith et al., 2009; Burn and Zhang, 2010), while deep snow cover acts as an effective
4 insulator, reducing heat loss and leading to warmer ground temperatures compared to areas with a thin
5 snow cover in winter (Burn et al., 2009). Permafrost distribution and behavior is also strongly linked to
6 vegetation characteristics as these influence the surface energy balance, snow accumulation, and soil
7 thermal properties (Burn, 2012). From a global perspective, permafrost-underlain watersheds,
8 particularly those at or near 0 °C, are extremely sensitive to warming as changes in ground thermal status
9 will alter all components of the hydrological cycle due to increases in subsurface storage for liquid water
10 (Woo, 2012). This also impacts forest ecosystems and function, leading to thaw-induced collapse of
11 forested peat plateaus and the loss of forest across vast parts of the boreal forest (Baltzer et al., 2014).

14 6.1-2 Permafrost Monitoring in Canada

16 ~~At its southern limits in Canada, permafrost—defined as ground that remains ≤ 0 °C for at least 2~~
17 ~~consecutive years—is sporadic and discontinuous, with annual temperatures near 0 °C, closely~~
18 ~~corresponding with the mean annual air temperature isotherm of 0 °C (Burn, 2012). There is a transition~~
19 ~~to continuous, deep, and colder permafrost with increasing latitude (Fig. 9).~~ Two of the largest monitoring
20 networks are maintained by the Geological Survey of Canada (GSC) and the Centre d'études nordiques,
21 Université Laval. The GSC, in collaboration with other partners, has been developing and maintaining a
22 network of sites that contribute to the Canadian Permafrost Monitoring Network as well as the
23 Circumpolar Active Layer Monitoring (CALM) programme and the Global Terrestrial Network for
24 Permafrost (GTN-P), providing long-term measurements of permafrost thermal state and active layer
25 conditions. About 75 sites operated by government and university scientists contribute to these
26 networks, with more than 20 boreholes available since the mid-1980s of up to 20 m depth in the
27 Mackenzie River valley and delta in the Northwest Territories (Fig. 9; Smith et al., 2005). A number of
28 additional monitoring sites were established during the International Polar Year (IPY) period 2007–2009,
29 bringing the total number of sites in North America up to 350 (Smith et al., 2010). The distribution of
30 these sites tends to be along roads and pipelines, while the boreholes themselves vary in depth,
31 measurement technique, and recording frequency; the network, however, provided a useful snapshot of
32 permafrost in northern North America for the IPY period (Smith et al., 2010). It is important to note that
33 for all ground temperature and active layer thickness data, local site characteristics and conditions can
34 have a considerable influence on the variability of permafrost conditions (e.g., Burgess and Smith, 2003;
35 Smith et al., 2009). Useful reviews of permafrost conditions and their recent changes over northern
36 Canada are provided by Smith (2011), Burn (2012), Derksen et al. (2012), and Bush et al. (2014).

38 6.2-3 Changes in Permafrost

40 Observations from monitoring sites across northern Canada have shown that ground temperatures to
41 depths of 10–15 m or more have been increasing in response to recent climate warming (Smith et al.,
42 2005, 2010). The magnitude of this warming has varied with latitude, generally exhibiting greater trends
43 in the northern continuous permafrost zone, and more limited temperature change in the southern
44 discontinuous and sporadic zones, where permafrost thaw has been widespread. Smith et al. (2005) and
45 Smith (2011) presented temporal variations in permafrost temperature from the Canadian Permafrost
46 Monitoring Network and the GTN-P. They showed that in the southern Mackenzie Valley, where
47 permafrost is warmer than -0.3 °C and only about 10 m thick, there has been no significant warming of
48 permafrost in the last few decades (less than 0.1 °C per decade). It was noted that the absence of trends

1 is probably due to phase change/latent heat effects restricting further warming. A similar pattern was
2 observed for warm permafrost in the Takhini River valley, southern Yukon Territory (Burn, 1998). Further
3 north in the central and northern Mackenzie region, warming of shallow permafrost of between 0.3 and
4 0.6 °C per decade has occurred since the mid-1980s (Smith et al., 2005; Smith, 2011), consistent with the
5 increasing air temperature trends over the same period (cf. Fig. 3). In the Mackenzie Delta area,
6 permafrost warming has been significant, but locally variable. For instance, Burn and Kokelj (2009)
7 compared permafrost temperatures here in the 1960s and early-1970s with those based on data collected
8 during 2003–2007. They showed that near-surface ground temperatures had increased over that time by
9 about 1 to 2 °C in the tundra uplands to the east of the delta, and by 0.5 to 1 °C in the delta itself south of
10 tree-line, where greater snow depth may have reduced the sensitivity to climate warming.

11
12 As ground temperatures have risen, there has also been an increase in the thaw depth of the seasonal
13 active layer across northern Canada, and local to widespread permafrost degradation in the discontinuous
14 permafrost zone (Smith, 2011). In many areas, the melting of the near-surface ground ice has resulted in
15 decreased stability and strength of the substrate, leading to surface subsidence and waterlogging of soils
16 (both of which further amplify permafrost thaw), thermokarst development, and collapse of forested peat
17 plateaus (Jorgenson et al., 2008; Smith et al., 2008; Baltzer et al., 2014). Burn and Kokelj (2009) reported
18 inter-annual variability of maximum seasonal thaw depths that ranged between 0.3 and 0.55 m, with a
19 trend toward increasing thaw depths of 0.08 m over the period 1983–2008 in the outer Mackenzie Delta
20 area. Smith et al. (2009) examined summer thaw depths over 100 × 100 m probing grids at various CALM
21 sites along the Mackenzie Valley, and found that intra-site variability can be high with substantial variation
22 in active layer depths where surface organic soils and moisture content are high and spatially variable.
23 Their 8-year (1998–2005) study period was too short to detect temporal trends, but they were able to
24 relate grid mean thaw depths to late summer thawing degree days on a site-by-site basis, and they showed
25 a latitudinal tendency of increasing maximum thaw depth southward. Burgess and Smith (2003)
26 demonstrated that rates of increasing thaw depth were generally greatest in the first 7–8 years after
27 disturbance at sites along a pipeline corridor between Norman Wells and northern Alberta. Permafrost
28 thaw had continued, albeit at a slower rate, for at least 17 years after 1985, reaching depths of 3–4 m in
29 colder lacustrine soils to over 7 m in coarse mineral soils. Further south, James et al. (2013) evaluated
30 changes in permafrost in the southern discontinuous permafrost zone at 55 sites along the Alaska Highway
31 from Fort St. John, BC to Whitehorse, YT, and observed that at almost half of the sites where permafrost
32 existed in 1964, it has since disappeared. Where permafrost remains, it has become patchy and thin (<15
33 m), has a greater active layer thickness, and has warmed to temperatures of between –0.5 and 0 °C. Their
34 results have indicated a northward shift in the limit of permafrost for this region.

35
36 Studies have examined changes in the spatial extent and connectivity of permafrost in the Taiga and
37 Boreal Plains within the discontinuous zone, and generally found that it is shrinking in area and becoming
38 increasingly fragmented. Beilman and Robinson (2003) used historic aerial photos and recent satellite
39 imagery to compare permafrost extent between about 1950 and 2000, and found that at peatland sites
40 in the Northwest Territories, as much as 50% of permafrost had degraded and thawed, while in more
41 southern locations in Alberta, Saskatchewan, and Manitoba, between 30% and 65% of localized
42 permafrost has thawed. Similar analysis at the Scotty Creek Research Basin near Ft. Simpson (Quinton et
43 al., 2011) showed that the extent of forest-covered permafrost plateaus had decreased by over 38% in
44 localized areas between 1947 and 2008, with associated collapse of black spruce forest on the plateaus
45 and an increasingly connected surface drainage network. This has the potential to substantially alter basin
46 runoff production in the region (Connon et al., 2014). Further, it has been shown that the rates of
47 permafrost and forested plateau loss have been accelerating, with the average rate of loss in 7 different

1 local areas of interest being over 3 times greater during the period 2000–2010 compared to 1977–2000
2 (Baltzer et al., 2014).

3 4 **7. Freshwater Ice Cover**

5 6 7.1 Context

7
8 Virtually all lake and river systems in the interior of western Canada are seasonally ice-covered, and this
9 influences the timing and severity of low flows and floods associated with freeze-up and break-up periods
10 (Beltaos and Prowse, 2009; Prowse, 2012) and has an important role in the ecology of these systems
11 (Prowse, 2001; Prowse and Culp, 2003). Spatially, ice thickness ranges from skims in more temperate
12 southern parts of this region to several meters in colder northern parts, while ice cover duration ranges
13 from being a transient feature to existing for over six months of the year (Prowse, 2012). The formation
14 and melt of the ice cover are sensitive to air temperature, precipitation, and other meteorological
15 variables, which influence the composition, thickness, and stability of the ice (Beltaos and Prowse, 2009).
16 In addition to thermal properties, river ice formation and break-up also depend on hydrodynamic
17 processes that affect the mechanical properties of the ice (Beltaos, 2003). Ice cover thickness, timing (i.e.
18 freeze-up and break-up), and duration therefore respond to changes in climatic conditions and other
19 controls such as changes in terrestrial hydrologic regimes (Prowse et al., 2011a, b). Some useful reviews
20 on the climatic controls, historical trends, and future projections of river ice formation and break-up are
21 provided by Prowse and Beltaos (2002), Beltaos and Burrell (2003), Prowse and Bonsal (2004), Prowse et
22 al. (2007), Beltaos and Prowse (2009), and Prowse (2012).

23 24 7.1.2 Ice Cover Monitoring in Canada

25
26 There are limited *in situ* records of ice cover variables over the interior of western Canada, which poses a
27 challenge for systematic regional monitoring and assessment of ice cover regimes and their changes over
28 time. Virtually all lake and river systems in the interior of western Canada are seasonally ice covered, with
29 maximum thickness ranging from skims in more temperate southern parts of this region to several meters
30 in colder northern parts, and ice cover duration ranging from being a transient feature to existing for over
31 six months of the year (Prowse, 2012). Various observations on freshwater and coastal sea ice conditions
32 in Canada have been gathered into a common database known as the Canadian Ice Database (CID) in
33 order to aid climate monitoring efforts and improve numerical prediction models and remote sensing
34 methods (Lenormand et al., 2002). The database contains records related to ice thickness, freeze-up and
35 break-up dates for 757 sites across Canada (including 312 lakes and 288 rivers). During the late-1980s
36 and 1990s, however, there was a drastic decline in the observation network and more recent observations
37 only represent a very small percentage of those available in the mid-1980s. Lenormand et al. (2002)
38 provide a complete description of the CID and its historical evolution. A volunteer monitoring program,
39 IceWatch (www.naturewatch.ca/icewatch/), began in 2001 and builds on the CID. Useful reviews of the
40 climatic controls, historical trends, and future projections of river ice formation and break-up are provided
41 by Prowse and Beltaos (2002), Beltaos and Burrell (2003), Prowse and Bonsal (2004), Prowse et al. (2007),
42 Beltaos and Prowse (2009), and Prowse (2012).

43 44 7.2.3 Changes in Ice Cover

45
46 Observations have shown a general reduction in ice cover duration on lakes and rivers across much of
47 Canada since the mid-20th century, due primarily to earlier spring break-up (Prowse, 2012). Zhang et al.
48 (2001a) and Burn and Hag Elnur (2002) analyzed trends in a number of hydro-climatic variables from the

1 Canadian Reference Hydrometric Basin Network (RHBN) database (see next section), and found that the
2 break-up of river ice and the spring freshet showed significant trends toward earlier occurrence at many
3 sites in Canada over the latter half of the 20th century. The most pronounced changes were observed in
4 western and southwestern Canada, with a greater degree of change over the last few decades of the
5 century. Lacroix et al. (2005) used observations from the CID to examine spatial trends in river freeze-up
6 and break-up over Canada and related these to the timing of the autumn and spring 0 °C isotherms. They
7 also found significant trends toward earlier spring break-up in this region over the latter part of the 20th
8 century, while autumn freeze-up patterns displayed greater regional variability and a mix of both
9 increasing and decreasing trends. Strong correlations were found between break-up dates and the spring
10 0 °C isotherms, but there were fewer significant associations between freeze-up and autumn 0 °C
11 isotherms. Similar patterns were found by Duguay et al. (2006) for lake ice observations as part of the
12 CID. Bonsal and Prowse (2003) showed that significant trends toward earlier spring 0 °C isotherm dates
13 have occurred over most of western Canada since 1950, whereas autumn isotherm dates showed little
14 change over most of Canada. Overall, the spatial and temporal trends in freeze-up and break-up closely
15 correspond to those of surface air temperature (Prowse, 2012).

16
17 The magnitude of trends in ice cover formation, break-up, and overall duration has varied among studies,
18 depending on datasets and temporal intervals of analysis. Lacroix et al. (2005) reported trends in river ice
19 break-up over Canada ranging from 1.0 to 2.2 days earlier per decade, and trends in freeze-up ranging
20 from 1 day later to 0.1 days earlier per decade, depending on the temporal analysis period (i.e., various
21 periods between 1950 and 1998). In the Mackenzie River basin, de Rham et al. (2008) found earlier trends
22 in spring break-up of about 1 day per decade in the upstream portions of the major tributaries over the
23 period 1970–2002. Using a combination of *in situ* measurements and remotely sensed observations to
24 examine lake ice phenological events, Latifovic and Pouliot (2007) found mostly earlier trends in lake ice
25 break-up that averaged –2.3 days per decade for the period 1970–2005, and mostly later trends in lake
26 ice formation that averaged 1.6 days per decade for the same period. It has been noted that based on
27 various northern cold region analyses, overall long-term increases in autumn and spring air temperatures
28 of 2–3 °C have been associated with a roughly 10-day delay in freeze-up and 15-day advance in break-up,
29 shortening the ice cover period by almost a month in some cases (Prowse and Bonsal, 2004; Prowse et
30 al., 2007).

31
32 Other characteristics of the seasonal ice cover are important, such as trends in the severity of river ice
33 break-up events and ice thickness, but there have been few broad-scale analyses in Canada. In the
34 Mackenzie Delta, a widespread tendency toward earlier break-up initiation and peak water level has been
35 reported, associated with a tendency towards the occurrence of more ice-driven break-up events as
36 opposed to discharge-driven events (Goulding et al., 2009). Ice-driven events are characterized by higher
37 upstream backwater levels from ice jams and earlier break-up initiation of a more competent ice cover.
38 It has been suggested that it is mid- to late-spring air temperatures and the pre-break-up melt and runoff
39 period that more strongly influence the timing of break-up than winter temperatures and maximum ice
40 thickness (Prowse and Bonsal, 2004). Prowse (2012) noted that the CID observations of Lenormand et al.
41 (2002) over Canada do not show any obvious trends in ice thickness over the latter part of the 20th century.

42

43 **8. River Discharge**

44

45 8.1 Context

46

47 The flow of water in rivers and streams provides a vital water resource to society and serves an important
48 ecological role; however, hydrological extremes represent serious potential risks and costs to society, such

1 as during floods or droughts. River discharge represents the spatially integrated response to inputs from
2 rainfall, snowmelt, and glacier melt, and its character is influenced by internal watershed storage capacity,
3 vegetation, soils, topography, and other landscape features affecting the delivery of water to the stream
4 channel network. In cold regions, the response to input events is highly sensitive to antecedent conditions
5 in the basin, including preconditioning of soils, snowpacks, surface and sub-surface water storage, and
6 surface drainage network connectivity (Spence and Woo, 2008; Carey et al., 2010; Bring et al., 2016), while
7 the nature of response is also dependent on the character of precipitation and the storm types (e.g., Shook
8 and Pomeroy, 2012). The timing, magnitude, duration, and other characteristics of discharge are
9 therefore closely linked with climate and other Earth system components, and are sensitive to their
10 changes within the watershed over multiple timescales.

11 12 *8.1-2 Canadian Reference Hydrometric Basin Network*

13
14 The Water Survey of Canada (WSC) maintains a network of over 2000 hydrometric monitoring stations
15 across Canada, and daily and monthly mean streamflow and stage levels are available through their
16 hydrometric database (HYDAT; www.ec.gc.ca/rhc-wsc/). A subset of the long-term WSC observing
17 stations was selected in the mid-1990s to characterize either pristine or stable hydrological conditions
18 without significant impacts from flow regulation or upstream diversions, and with good quality records
19 for at least 20 or more years (Harvey et al., 1999). This subset comprises the Reference Hydrometric Basin
20 Network (RHBN) in Canada, presently consisting of 217 active streamflow stations across the country with
21 an average record length of over 50 years. The RHBN streamflow records are considered suitable for
22 climate related studies while most other non-RHBN sites are likely to be affected by other drivers
23 (Personal communication, P. Whitfield, April, 2014). Zhang et al. (2001a) and Burn and Hag Elnur (2002)
24 provide detailed descriptions of the RHBN and the selection criteria used to include stations. In particular,
25 a high level of accuracy was ensured by assessing the quality of data for each station with regard to the
26 reliability of the stage–discharge relationship, stability of channel geometry, and reliability of water level
27 and discharge measurement during ice-covered conditions (Zhang et al., 2001a). Whitfield et al. (2012)
28 note that the RHBN records have been used in over 25 studies since the network was established, and
29 they stress the importance of maintaining such reference hydrologic networks of long-term, quality time
30 series data in relatively undisturbed regions.

31 32 *8.2-3 Changes in River Discharge*

33
34 Many studies have examined variability and trends in the magnitude, timing, and other characteristics of
35 river discharge over Canada (e.g., see Koshida et al., 2015; Mortsch et al., 2015). Direct comparison of the
36 results is often difficult due to differences in temporal analysis periods, statistical methodology, region of
37 focus, and datasets (i.e., RHBN and non-RHBN). Annual mean flow has been observed to vary regionally,
38 with studies documenting both increasing and decreasing trends since the 1960s (Prowse et al., 2009b).
39 Major river systems such as the Mackenzie and Nelson (which includes the Saskatchewan River system)
40 have shown no detectable long-term trends at their mouths over this time (Woo and Thorne, 2003; Déry
41 and Wood, 2005; McClelland et al., 2006; Déry et al., 2011), while statistically significant declines in annual
42 flow have been observed for some smaller systems within these, such as the Athabasca River and its
43 tributaries, and other rivers draining from the eastern slopes of the southern Rocky Mountains (Burn et
44 al., 2004b; Rood et al., 2005; St. Jacques et al., 2010; Peters et al., 2013). Seasonally, a consistent pattern
45 of increasing flow in the winter has been reported—especially for the Mackenzie River and many of its
46 northern tributaries—with significant trends in the annual minimum discharge and lower flow percentiles
47 (Burn et al., 2004b; Rood et al., 2008; St. Jacques and Sauchyn, 2009). Discharge rates during other times
48 of the year have varied, but studies across the region have found decreasing trends in the annual

1 maximum discharge rate and spring freshet flow volume (Burn et al., 2004b; 2008; 2010). Burn et al.
2 (2008; 2010) have also reported increasing trends in the number of rainfall–runoff events later in summer
3 and their peak magnitude, but little trend in runoff volume of these events.
4

5 To synthesize recent changes in flow across western Canada, we performed an analysis of monthly and
6 annual discharge from RHBN records. Trends were computed for 3 different periods beginning in 1960,
7 1970, and 1980, and all ending in 2010. The analysis was restricted to currently active stations, and
8 excluded most higher order rivers as these are mostly affected by upstream regulation and other human
9 influences (and are thus not part of the RHBN) while the RHBN includes only stations with no upstream
10 flow regulation; this eliminates many higher order rivers, as few are unaffected by regulation. Missing
11 data thresholds were set at no more than 3 days per month and 4 years in the analysis period, and records
12 with missing data in excess of these limits were rejected. The Mann–Kendall rank trend test (Mann, 1945;
13 Kendall, 1975) was used, with significance assessed at the 95% confidence level, and trend slope was
14 estimated based on the method of Sen (1968).
15

16 Annual and January trend results are shown in Fig. 10. Trends since 1960 are not shown as most were
17 rejected due to missing data. Annual flows exhibited a mixture of increasing and decreasing trends (Fig
18 10a), while their magnitude, significance, and in some cases, direction, depended on the length of the
19 analysis period. There is no well-defined regional pattern of consistent, long-term annual discharge
20 change, except perhaps in the southern mountain regions of British Columbia and Alberta, where flows
21 at a number of stations have decreased slightly over time. During the winter months there was a clear
22 pattern of increasing flows across the North, with many stations exhibiting significant positive trends in
23 all 3 periods (January flows in Fig 10b provide a good example). Due to the low flow rates at this time of
24 year, however, this has little influence on the annual mean flow of most rivers. In the southern half of the
25 domain, trends in winter and spring discharge showed no consistent regional pattern, except later during
26 summer and fall (not shown), when many streams and rivers in the North Saskatchewan and Peace–
27 Athabasca River basins exhibited significant declining flows.
28

29 The most consistently reported characteristic of discharge trends across western Canada has been an
30 earlier onset of the spring freshet since at least the mid-1960s, as indicated by features such as the initial
31 hydrograph rise and the timing of peak spring flow (Woo and Thorne, 2003; Burn et al., 2004a; 2004b;
32 2008; 2010; Burn, 2008; Abdul Aziz and Burn, 2006; Rood et al., 2008; Cunderlik and Ouarda, 2009). Woo
33 and Thorne (2003) reported that between 1973 and 1999 the date of spring hydrograph rise for the
34 Mackenzie River and several of its major tributaries (i.e. the Peace, Liard, and Slave Rivers) advanced by
35 about three days per decade. Burn (2008) examined the trend behavior of 9 measures of the timing of
36 runoff for gauging stations within the Liard, Peace, and Athabasca River basins over various periods during
37 the late-20th century. They found field significant trends (i.e., a collection of time series for a hydrologic
38 variable that exhibits a greater number of trends than expected by chance) in several of the variables
39 analyzed, and in particular they observed strong shifts in timing towards earlier spring freshet for all
40 periods, especially in headwater catchments. Similar results were obtained in the Prairies (Burn et al.,
41 2008) and the Rocky Mountain eastern slopes (Rood et al., 2008). These trends are consistent with
42 increasing spring air temperatures and an earlier occurrence of snowmelt reported across the domain
43 (Sections 2 and 4).
44

45 The timing of low flows has also received attention (Ehsanzadeh and Adamowski, 2007; 2010; Khaliq et
46 al., 2008; Burn et al., 2010). Annual low flow events of various durations have generally been observed
47 to show a decreasing (i.e. arriving later) trend in southwestern Canada and an increasing trend in the
48 northwest (Khaliq et al., 2008). In the prairie region, Burn et al. (2008) found a decreasing trend in the

1 timing of summer rainfall-driven peak flow events, while over a broader part of western Canada, Cunderlik
2 and Ouarda (2009) found no significant trends in the timing of fall rainfall-dominated high flow events.

3
4 Given the limited record length for flow analyses, the potential for computed trends to be influenced by
5 shorter term variability and large-scale modes of oceanic–atmospheric circulation variability is high. Woo
6 et al. (2006) discuss this issue and illustrate the difficulty in attributing trends detected in short records to
7 long-term climate change. Some studies have accounted for modes of large-scale variability to examine
8 the trends in the absence of their influence. For example, St. Jacques et al. (2010) removed the signal due
9 to the PDO in flows at a number of stations across the southern Canadian Rockies and southern Alberta,
10 and found that observed and naturalized flows still exhibited predominantly decreasing trends since as
11 far back as the early 20th century. Burn’s (2008) analysis of flow timing indices in the Mackenzie River
12 basin examined the relationship with meteorological variables and 6 large-scale circulation indices. They
13 concluded that although several of the timing measures are related to large-scale circulation patterns,
14 there is a relationship with increasing spring air temperatures, and the large-scale periodic influences are
15 not a dominant contributor to the observed trends. Bonsal and Shabbar (2008) review the past Canadian
16 research on the impacts of large-scale circulation patterns on low flows, and report a higher frequency of
17 low flow events in western Canada associated with the warmer and drier conditions during El Niño events
18 and positive phases of the PDO and the PNA pattern. They also note that the spatial and temporal aspects
19 of the relationships between low flows and these climatic patterns are strongly influenced by local hydro-
20 climatic complexities, particularly in the mountainous watersheds.

21 22 **9. Discussion and Conclusions**

23 24 *9.1 Dataset Quality and Length*

25
26 The observational datasets described throughout this paper provide the best available long-term, quality
27 controlled products for examining Earth system change and variability in western Canada. Considerable
28 effort has gone into ensuring their reliability for trend detection and their intercomparability for regional
29 assessments, including corrections for known sources of inhomogeneity, station relocation, measurement
30 error, and data gaps. The data are, however, subject to several notable limitations. First, the hydro-
31 climatic observational network is sparse in northern Canada and regional assessments require
32 extrapolation and interpolation over vast areas. This is likely less problematic for air temperature than
33 for precipitation, for example, which can exhibit greater local variability and is more sensitive to the
34 density of the observing network and methods for adjusting systematic measurement errors (Rapačić et
35 al., 2015). Rapačić et al. (2015) noted that considerable care is needed when using gridded climate datasets
36 in local or regional scale applications in data sparse regions such as the Canadian Arctic, where, for
37 example, they found that the adjustments used in the CANGRD product led to a doubling of precipitation
38 trend magnitude compared to a multi-dataset average. Sparseness of the observing network is also an
39 issue in mountainous areas, where there are relatively few long-term sites at high elevations—a problem
40 that exists globally (Adam et al., 2006). Another challenge is mMonitoring in cold, harsh conditions and
41 the associated measurement error also presents a challenge; in particular, solid precipitation
42 measurements are highly susceptible to error as a result of wind and turbulence around the gauge and
43 require careful correction (Yang et al., 2005).

44
45 Another issue involves the record length of these observational datasets. The longest climate station
46 records go back to the early 20th century or late 19th century for a few select sites, while across much of
47 the northern parts of western Canada, there are very few climate records going back much further than
48 about 1950. This is also a problem for other variables, such as snow and freshwater ice cover, and

1 particularly river discharge. As noted previously, it is difficult to discern long-term trends from periodic
2 effects in short records, and the results may be highly sensitive to analysis period. Large-scale oceanic–
3 atmospheric circulation patterns, such as ENSO and PDO, are known to influence the variability of
4 cryospheric and hydrological systems on time-scales of years to decades (Whitfield et al., 2010), and can
5 thus significantly complicate the detection and attribution of trends (Woo et al., 2006). In many instances,
6 the closure of stations or the automation of measurement procedures has caused gaps or inconsistencies
7 in the data, also limiting the length of observational datasets. Since about the mid-1990s there has been
8 a considerable reduction in active climate, snow, and discharge monitoring stations across this region,
9 which restricts the analyses of long-term change.

10 11 *9.2 Climatic, Cryospheric, and Hydrological Change at the WECC Observatories*

12
13 Despite the data limitations noted, observations have shown clear and systematic patterns of change in
14 climatic regime and cryospheric response over western Canada. The various lines of evidence are
15 consistent and mutually supportive. Warming has been pervasive, especially during winter and spring and
16 at higher latitudes (warming rates in western Canada are among the highest globally), while changes in
17 precipitation have been more varied, both regionally and seasonally. Widespread decreases in winter
18 precipitation in much of southwestern Canada have been observed, and a decline in the fraction of
19 precipitation falling as snow is very likely associated with rising winter/spring temperatures and a shift in
20 the timing of the 0 °C isotherm. The snow to rain transition is sensitive to increasing temperature,
21 particularly when and where precipitation occurs at temperatures near 0 °C (Rasouli et al., 2015). These
22 changes are also driving widespread reductions in snow depth, snow cover extent and duration, and
23 freshwater ice cover, primarily in spring as opposed to fall. This has led to an earlier occurrence of the
24 spring freshet across the region. Warmer air temperatures are associated with rising permafrost
25 temperatures, thawing and degradation of permafrost, and increasing glacier melt. Recent declines in
26 winter and annual glacier mass balance at several sites have been attributed to reduced snow
27 accumulation in association with a shift in the PDO in 1976 (Demuth and Keller, 2006; Moore et al., 2009),
28 but may also be due in part to a shift in precipitation phase from snow to rain. Conditions observed in the
29 winter and spring of 2015 across the southern Canadian Rocky Mountains may be indicative of what the
30 future holds under an increasingly warmer climate. Anomalously warm conditions led to reduced
31 snowpacks and early disappearance of snow cover (despite near-normal winter precipitation along the
32 central ranges), also leading to end-of-summer snowlines retreating off the tops of most glaciers and
33 record net negative glacier mass balance for the period of available measurements (personal
34 communication, M. Demuth, 2015; Demuth and Ednie, 2016).

35
36 In general, river discharge magnitude has not shown long-term and regionally coherent trends, other than
37 increasing winter flows in the north. This may partly reflect aspects of periodicity and the influence of
38 large-scale modes of climate variability over inter-annual to inter-decadal scales (e.g., ENSO, PDO, etc.).
39 However, it has been found that these modes are not the sole driver of observed trends in other variables
40 (Vincent et al., 2015). The mixed responses in discharge magnitude are probably the result of interactions
41 between multiple processes and drivers—which may be confounding and multi-directional—across
42 various temporal and spatial scales and geographic locations.

43 44 *9.3 Insights on Change from the WECC Observatories*

45
46 While in most cases, the length of record at the CCRN WECC observatories (Fig. 1) precludes robust-time
47 series analysis as conducted on the national climatological and hydrometric data sets, the observatories

1 provide critical platforms to understand hydrological responses and process interactions, and to diagnose
2 how changes are manifested at scales where field-based scientific investigations are undertaken.

3
4 One of the most notable changes in the cryosphere is the widespread thaw of permafrost across the
5 circumpolar region (Hinzman et al., 2005; Schuur et al., 2015). ~~From a global perspective, permafrost~~
6 ~~underlain watersheds, particularly those at or near 0°C are extremely sensitive to warming as changes in~~
7 ~~ground thermal status will alter all components of the hydrological cycle due to increases in subsurface~~
8 ~~storage for liquid water.~~ In the Mackenzie and Yukon River basins of North America, recent changes in
9 winter low flows have been ascribed to permafrost thaw and adjusting pathways (Walvoord and Striegl,
10 2007; St. Jacques and Sauchyn, 2009). In contrast, at the headwater and meso-scales (<1000 km²), these
11 trends are rarely observed, ~~creating some uncertainty as to whether activation of new and deeper~~
12 ~~subsurface flow pathways is the cause of change at larger scales and it is equivocal as to whether~~
13 ~~observations at larger scales are directly reactivation of subsurface flowpaths.~~ Other factors such as
14 increased rainfall (Spence and Rausch, 2005), changing seasonality of precipitation (Whitfield et al., 2004),
15 changing landscape connectivity (Connon et al., 2014), thermokarst development (Kokelj et al., 2013;
16 Malone et al., 2013), and over-winter thaw events (Spence et al., 2014) act to alter streamflow volume
17 independently of widespread permafrost thaw.

18
19 Several WECC observatories have been operational since the 1990s in permafrost-underlain catchments
20 and have undergone change associated with warming. The most dramatic change has been observed at
21 Scotty Creek in the Taiga Plains near Fort Simpson, NT, where degradation in permafrost has resulted in
22 decreasing and fragmented forest cover (Baltzer et al., 2014), and increasing wetland coverage and
23 connectivity and drainage efficiency (Quinton et al., 2011; Connon et al., 2014). From 1947 to 2008,
24 permafrost coverage declined at the site from 70% to 43% (Chasmer et al., 2010) and wetland permafrost-
25 free areas expanded and coalesced. Connon et al., (2014) showed that increases in flows observed in the
26 region were not a result of reactivation of subsurface flow pathways from permafrost thaw, but from an
27 increased connectivity of surface pathways, and to a lesser extent increases in overall precipitation. Bogs
28 that were formerly isolated on the landscape became incorporated into the basin drainage network,
29 increasing runoff efficiency through time.

30
31 Further north at the Baker Creek, NT, WECC observatory, changing precipitation regimes from warmer
32 temperatures are more strongly influencing the hydrological response of this watershed. Here, the
33 Precambrian Shield geological setting, abundant lakes, and thin glacial-fluvial soils are not particularly
34 sensitive to changes in thermal regime, but changes in lake storage can rapidly change the runoff
35 contributing area available for river discharge formation. The WECC observatory has a distinct threshold-
36 response runoff regime (Spence, 2006) where increasingly large areas of the watershed connect and can
37 contribute to runoff as lake storage increases (Phillips et al., 2011). Spence et al. (2011) have documented
38 sudden changes in streamflow regimes as the climate has changed. Previously, peak flows were observed
39 during spring freshet in response to the relatively rapid snowmelt, whereas in more recent years, large
40 summer and fall precipitation events that are rare in the historical record have resulted in enhanced late-
41 season flows. In addition, Spence et al., (2015) revealed the emergence of new streamflow regimes as a
42 result of late-season rains, with enhanced winter storage and streamflow being characteristic of a
43 transition from a nival toward a pluvial regime, and they showed how this influences the biogeochemical
44 system.

45
46 Other WECC observatories underlain with permafrost (notably Wolf Creek, YT, and Trail Valley Creek, NT)
47 have undergone less dramatic changes since the 1990s despite warming air temperatures, although
48 important insights to potential response from change are emerging. In Trail Valley Creek where

1 permafrost is continuous, active layers have increased and there have been gradual expansions of taller
2 shrub vegetation (Lantz et al., 2012). Shi et al. (2015) examined several timing metrics of the runoff
3 hydrograph and revealed complex responses for 27 years of data. Unlike what has been reported in
4 southern latitudes (Berghuijs et al., 2014), a decline in total winter snow accumulation had only minor
5 influence on the timing of springtime streamflow. Offsetting effects of springtime temperature
6 fluctuations and winter warming resulted in small changes to the freshet hydrograph, whereas delayed
7 springtime rain had a more notable impact on the late-summer flow regimes. Overall, despite major
8 changes in climate, vegetation, and active layer thickness, changes to streamflow were limited. At Wolf
9 Creek, an alpine watershed underlain with discontinuous permafrost, recent changes in shrub abundance
10 at higher elevation has resulted in some changes in accumulation of snow (Pomeroy et al., 2006), yet
11 responses in the streamflow hydrograph to change have yet to be realized with 18 years of data. Using a
12 hydrological model to simulate future climates, Rasouli et al., (2014) showed that while the hydrological
13 impacts of climate warming are unequivocal (i.e., reduced snow contributions to streamflow, shorter
14 snow-covered period, greater evapotranspiration), the magnitude and direction of the impact of warming
15 on streamflow depends more on changes to future precipitation than on future warming.

16
17 In more southerly locations, the influence of land use and climate change combined to dramatically alter
18 hydrological regimes in Smith Creek, MB, an agricultural basin characteristic of the southeastern prairie.
19 Warming temperatures, increased rainfall fraction, earlier melt, and more multiple-day rainfall events
20 have, in concert with extensive wetland drainage and increase in basin hydraulic connectivity, resulted in
21 an overall dramatic increase in flows generated from snowmelt, rain-on-snowmelt, and rainfall runoff
22 processes, with the greatest increases for rainfall runoff and a relative decline in the proportion of
23 streamflow derived from snowmelt from over 85% in the 20th century to less than 50% in the last five
24 years (Dumanski et al., 2015). Conversely, in the eastern slopes of the Rocky Mountains, Marmot Creek,
25 AB has shown considerable resilience in its flow regime. Despite winter warming, declining low elevation
26 snowpacks, forest cover reduction, and more recently, extreme weather, streamflow volume, timing, and
27 magnitude of peak have not changed since 1962 (Harder et al., 2015). An increase in the fraction of
28 precipitation occurring in spring and the frequency of multiple day rainfall events may have counteracted
29 warmer conditions and lower snowpacks at low elevations. Substantial sub-surface storage in the basin
30 may further buffer the impacts of climate variability. The devastating flood of 2013 that affected Calgary
31 and much of the eastern slopes was in fact buffered by processes operating within headwater catchments
32 as flow responses were less than anticipated based on precipitation statistics. Of particular interest is that
33 two years later in 2015, despite relatively normal total volumes of winter precipitation, smaller snowpacks
34 (due to more mid-winter rainfall and melt events) have resulted in dramatically lower streamflows and
35 record low groundwater levels, exacerbating the ecological and societal impact of drought conditions that
36 arrived in the spring.

37
38 Ongoing research at the WECC observatories will continue to address the interplay among changes in
39 climate, cryosphere, ecology, and the hydrological responses observed, and place larger-scale patterns in
40 a local context. There are a number of outstanding research questions that remain to be addressed and
41 will be a focus of ongoing research including: 1) the role of changing vegetation and ecological
42 assemblages (including land-use change) on hydrology at the WECC observatories, 2) the importance of
43 fall and over-winter hydrological processes, which are often neglected in field-based research yet are
44 increasingly being identified as important when evaluating change, 3) the influence of thawing permafrost
45 and a reduction in seasonally frozen ground on hydrological response across a gradient of watersheds in
46 different thermal and physiographic environments, and 4) how the decline in snow (in terms of volume,
47 duration, and as a fraction of total precipitation) will influence all aspects of the hydrological cycle. An
48 important activity of CCRN will be to develop conceptual and numerical models of change for each WECC

1 observatory using future climate scenarios and a common experimental framework linking all biophysical
2 aspects of these systems. While large national observation networks are critical for baseline assessment
3 of change, linking this to research at small scales is still required to interpret patterns observed in larger
4 basins and will be a focus of the network.

5 6 9.34. Concluding Remarks

7
8 ~~Despite the data limitations noted, observations have shown clear and systematic patterns of change in~~
9 ~~climatic regime and cryospheric response over western Canada. The various lines of evidence are~~
10 ~~consistent and mutually supportive. Warming has been pervasive, especially during winter and spring and~~
11 ~~at higher latitudes (warming rates in western Canada are among the highest globally), while changes in~~
12 ~~precipitation have been more varied, both regionally and seasonally. Widespread decreases in winter~~
13 ~~precipitation in much of southwestern Canada have been observed, and a decline in the fraction of~~
14 ~~precipitation falling as snow is very likely associated with rising winter/spring temperatures and a shift in~~
15 ~~the timing of the 0 °C isotherm. The snow to rain transition is sensitive to increasing temperature,~~
16 ~~particularly when and where precipitation occurs at temperatures near 0 °C (Rasouli et al., 2015). These~~
17 ~~changes are also driving widespread reductions in snow depth, snow cover extent and duration, and~~
18 ~~freshwater ice cover, primarily in spring as opposed to fall. This has led to an earlier occurrence of the~~
19 ~~spring freshet across the region. Warmer air temperatures are associated with rising permafrost~~
20 ~~temperatures, thawing and degradation of permafrost, and increasing glacier melt. Recent declines in~~
21 ~~winter and annual glacier mass balance at several sites have been attributed to reduced snow~~
22 ~~accumulation in association with a shift in the PDO in 1976 (Demuth and Keller, 2006; Moore et al., 2009),~~
23 ~~but are also likely due in part to a shift in precipitation phase from snow to rain. Conditions observed in~~
24 ~~the winter and spring of 2015 across the southern Canadian Rocky Mountains may be indicative of what~~
25 ~~the future holds under an increasingly warmer climate. Anomalously warm conditions led to reduced~~
26 ~~snowpacks and early disappearance of snow cover (despite near normal winter precipitation along the~~
27 ~~central ranges), also leading to end-of-summer snowlines retreating off the tops of most glaciers and~~
28 ~~record net negative glacier mass balance for the period of available measurements (personal~~
29 ~~communication, M. Demuth, 2015).~~

30
31 ~~In general, river discharge magnitude has not shown long term and regionally coherent trends, other than~~
32 ~~increasing winter flows in the North. This may partly reflect aspects of periodicity and the influence of~~
33 ~~large-scale modes of climate variability over inter-annual to inter-decadal scales (e.g., ENSO, PDO, etc.).~~
34 ~~However, it has been found that these modes are not the dominant driver of observed trends in other~~
35 ~~variables (Vincent et al., 2015). The mixed responses in discharge magnitude are probably the result of~~
36 ~~interactions between multiple processes which may be confounding and multi-directional across~~
37 ~~various temporal and spatial scales. Different drivers have acted across the region considered here; for~~
38 ~~instance, late season precipitation and permafrost decay have altered runoff regimes in the North, while~~
39 ~~an increase in the fraction of precipitation as rain and more multiple-day storm events have influenced~~
40 ~~flow generation in the South. Human influences also may affect the observed trends and in some~~
41 ~~instances, explain differences between RHBN and other flow records.~~

42
43 ~~To conclude we note that rapid warming is occurring throughout the interior of western Canada, as in~~
44 ~~other cold region environments globally, with profound implications in terms of landscape and~~
45 ~~hydrological and change. However, hydrological responses are complex and spatially variable, reflecting~~
46 ~~multiple process interactions at sub-basin scale that vary across the region, and depend strongly on~~
47 ~~changes in precipitation form and timing. There has been much speculation on the future trajectory of~~
48 ~~hydrological change in cold regions in relation to climate projections, but this has not often been grounded~~

1 on appropriate conceptual understanding and process-based representation of the underlying response
2 mechanisms. Given the complexities involved, this is required to develop plausible scenarios of change
3 for the 21st century, especially considering that while certain processes and their interaction are inherently
4 accounted for in some physically-based models, other factors are likely not (e.g., landcover changes such
5 as deglaciation, permafrost degradation, and changing vegetation and ecological assemblages, and their
6 influence on hydrology). Simple projections of change are therefore likely to be misleading, ~~predictions~~
7 ~~of change must be based on appropriate conceptual understanding and process-based modelling.~~ There
8 is clearly a need to diagnose and better understand the causes and mechanisms underlying the observed
9 pattern in flow regime and its variability over the region if projections of future change (and associated
10 societal impacts) are to be made with any confidence.

11 **Data Availability**

12
13 All datasets used here are publically available and can be accessed through the links and references we
14 have provided.
15

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17
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24
25

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Figures

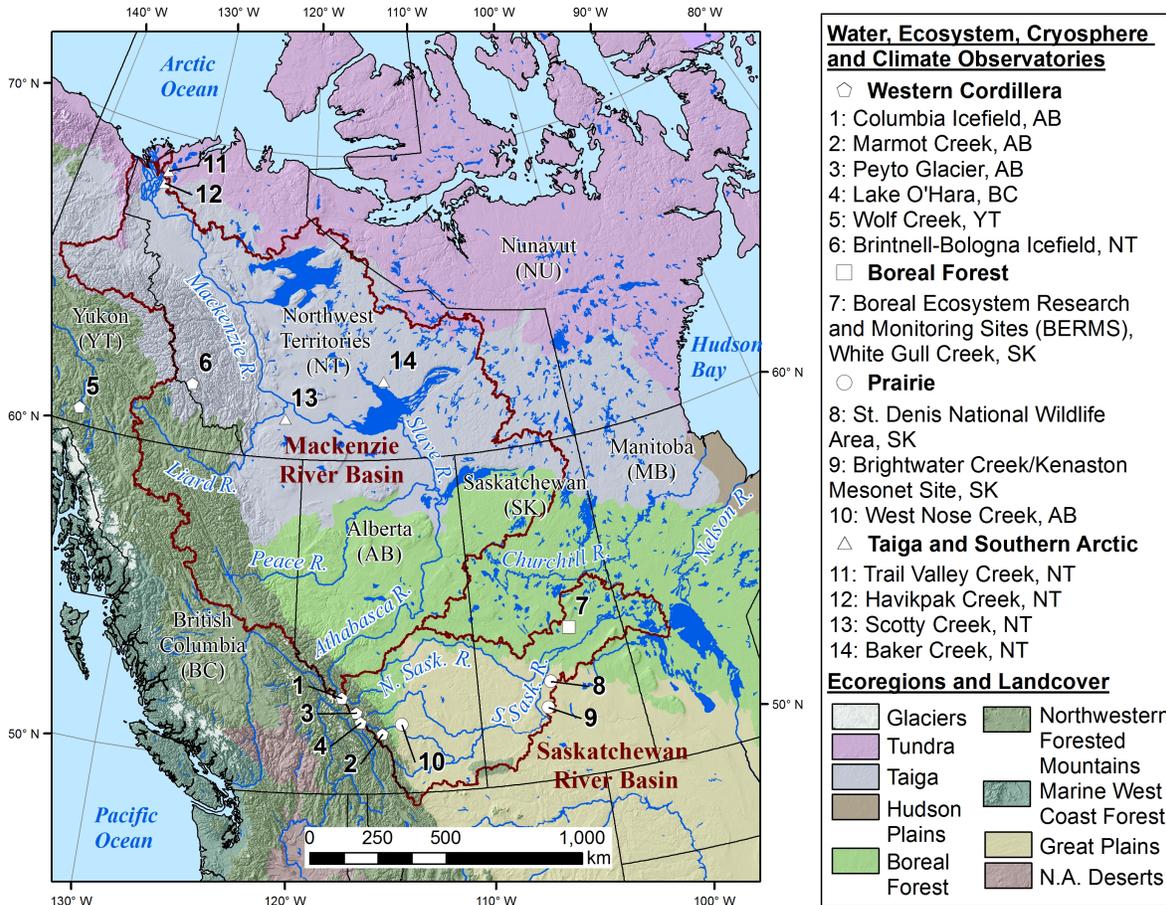


Figure 1. Map of the CCRN geographic study domain in the interior of western and northern Canada, showing the major river systems, ecoregions, and landcover, as well as the location of the WECC observatories. Source data is from the North American Environmental Atlas (<http://www.cec.org/naatlas/>) and the National Hydro Network (<http://www.geobase.ca>); the projection is UTM Zone 11 on the North American Datum of 1983.

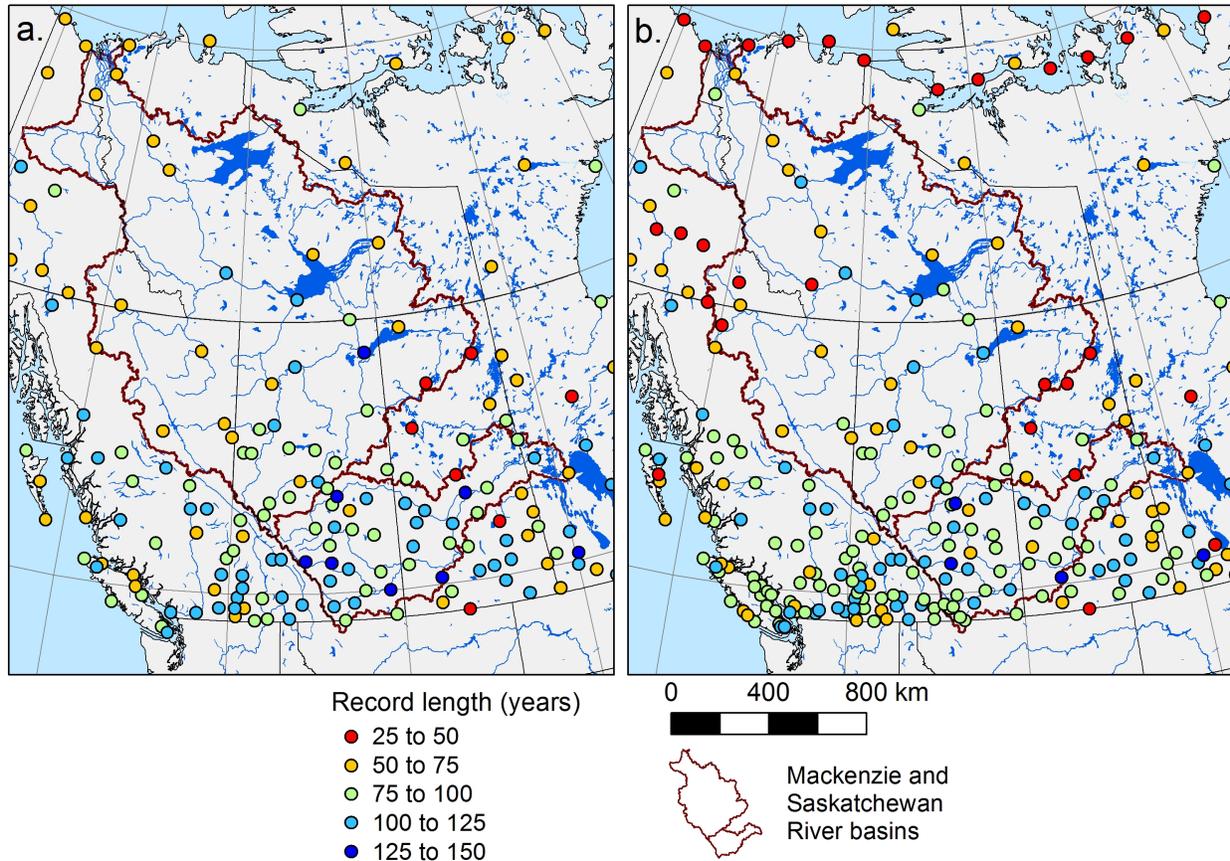


Figure 2. Map of station density and record length for a) temperature and b) precipitation measurements used in the AHCCD of Environment Canada (<http://www.ec.gc.ca/dccha-ahccd/>).

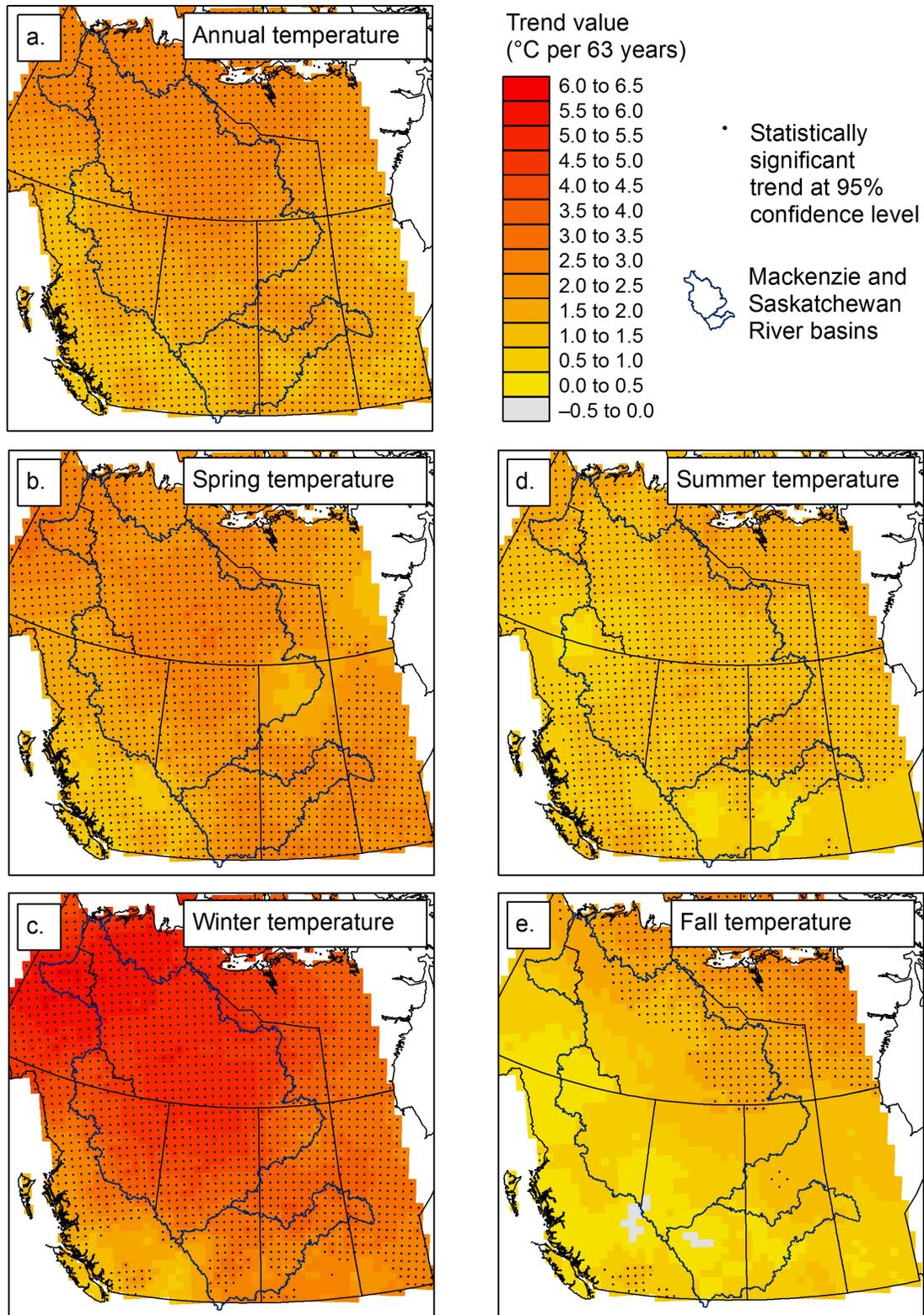


Figure 3. Spatial patterns of trends (°C per 63 years) in annual and seasonal average air temperatures over the period 1950–2012 across western Canada, based on analysis of CANGRD temperature data.

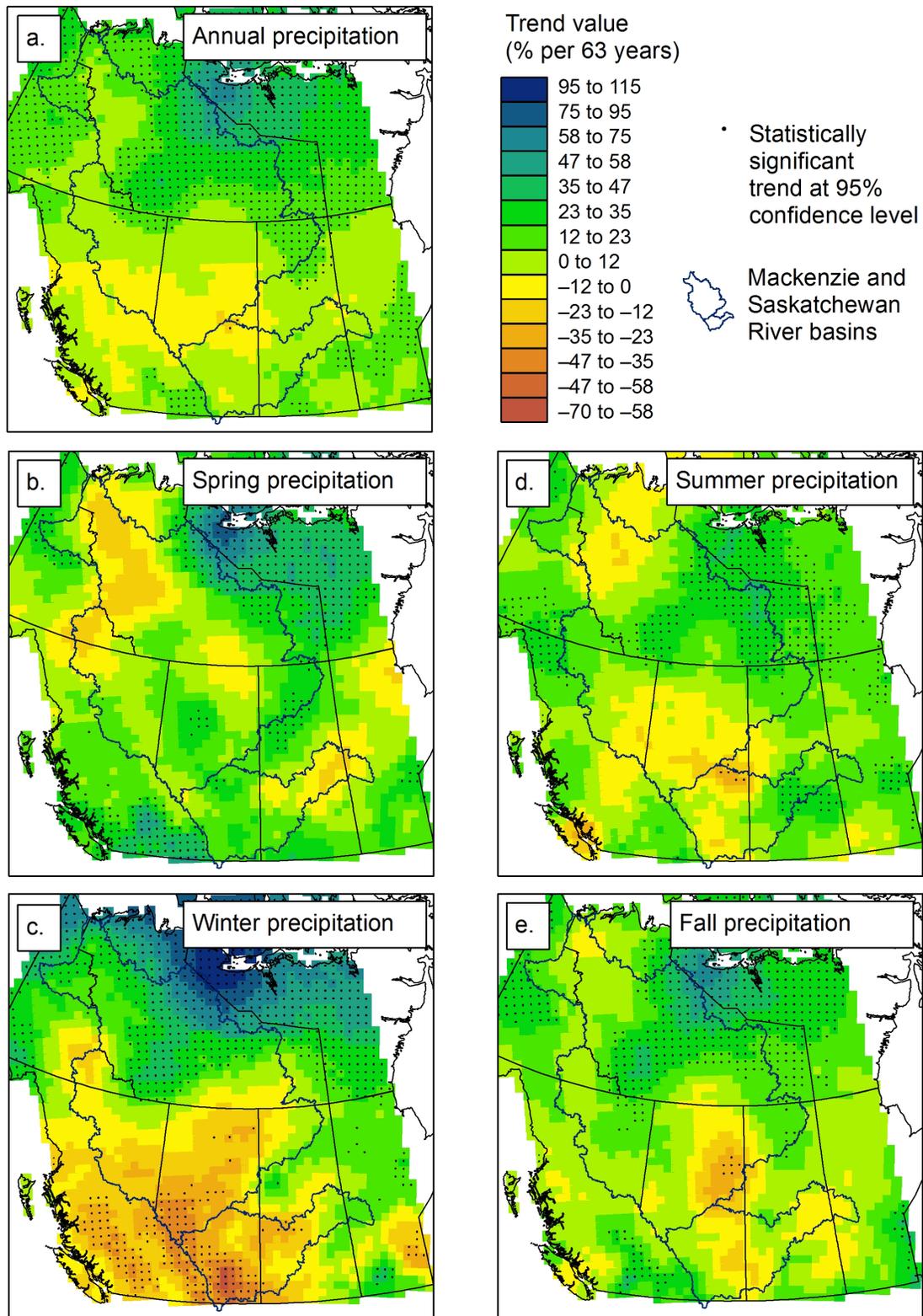


Figure 4. Spatial patterns of trends (percent per 63 years) in annual and seasonal totals of precipitation over the period 1950–2012 across western Canada, based on analysis of CANGRD precipitation data.

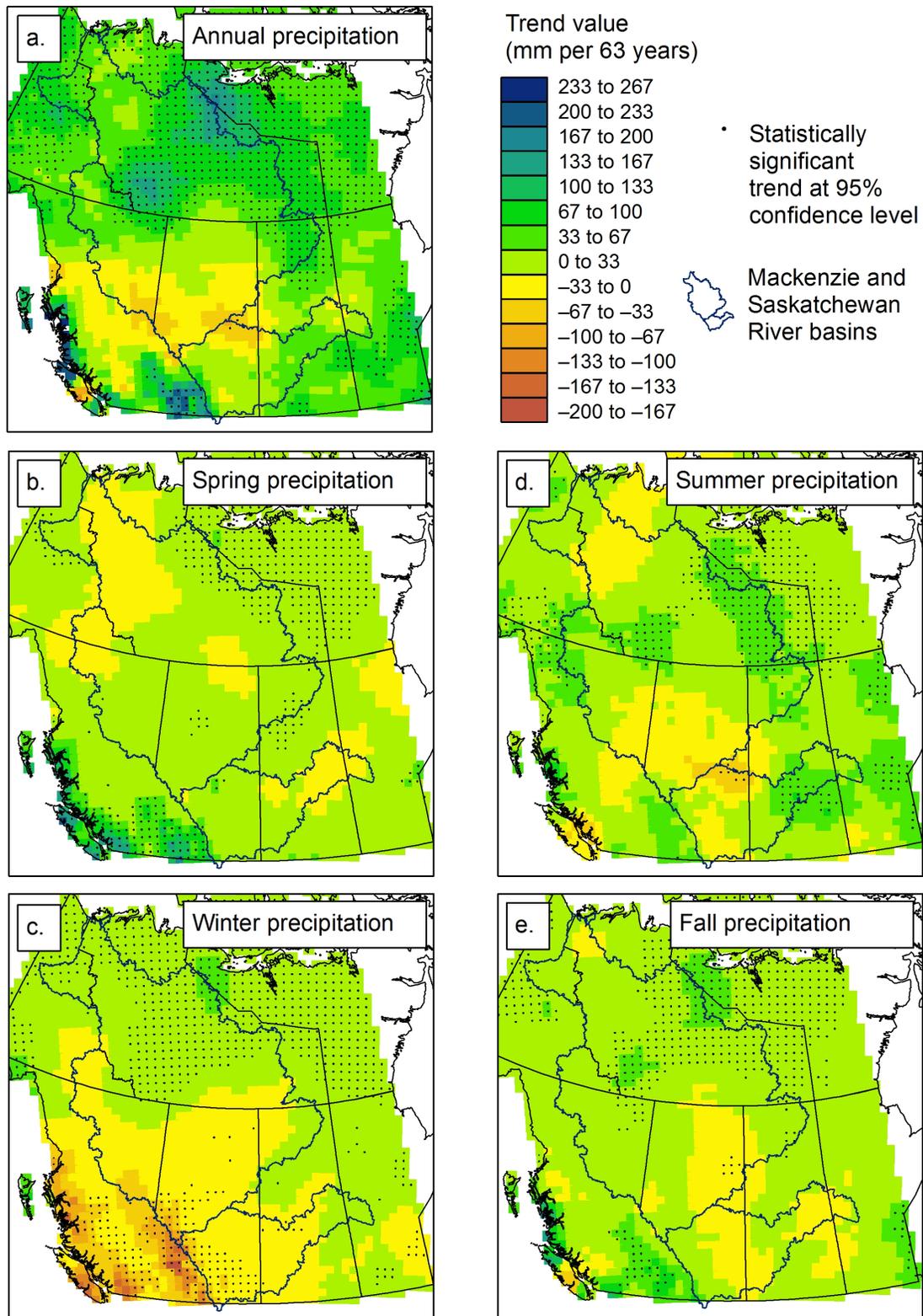


Figure 5. Spatial patterns of trends (mm per 63 years) in annual and seasonal totals of precipitation over the period 1950–2012 across western Canada, based on analysis of CANGRD precipitation data.

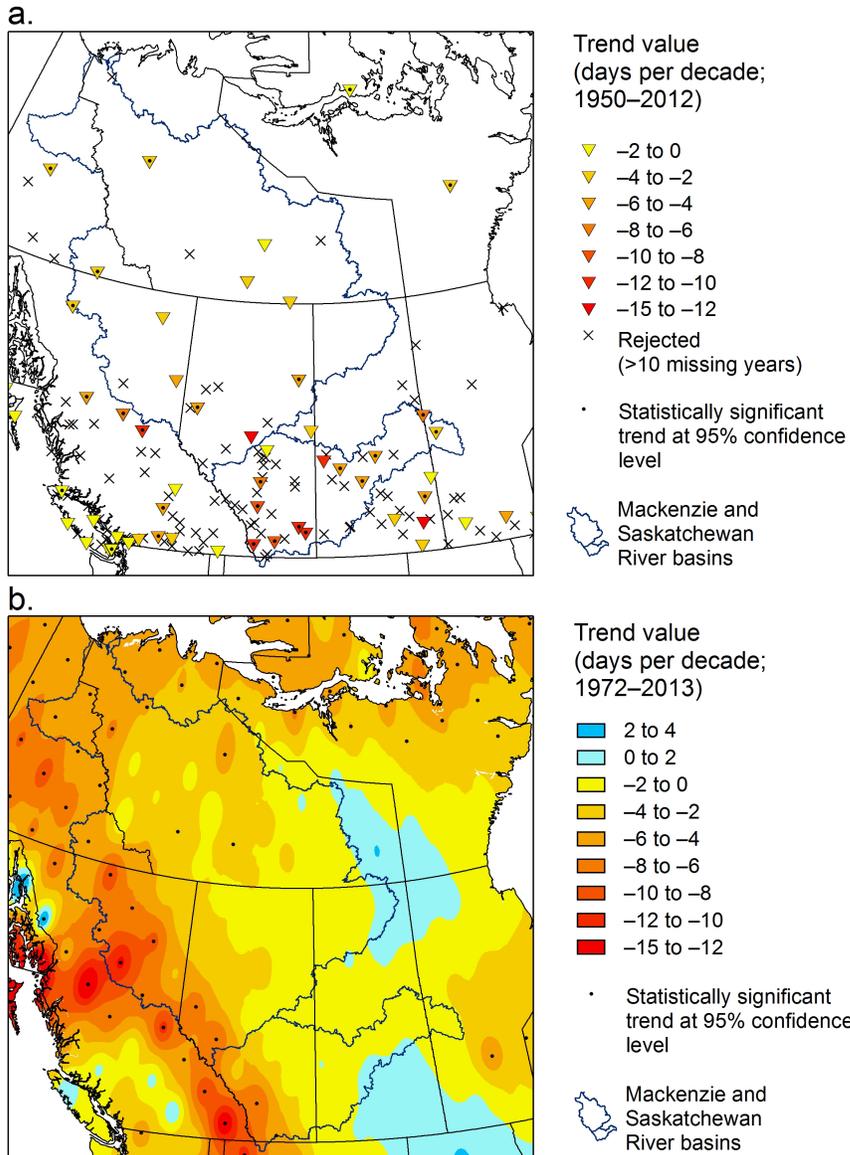


Figure 6. a) Trends in annual snow cover duration for the period 1950–2012 based on surface observations as part of the Canadian daily snow depth dataset. Stations were rejected in cases where the amount of missing data exceeded 10 years. b) Trends in spring season (February to August) snow cover duration for the period 1972–2013 based on the NOAA weekly snow cover product from the Rutgers University dataset (Rutgers data documentation at <http://climate.rutgers.edu/snowcover/docs.php?target=vis>). In b, the spatial patterns are based on inverse distance weighting of the points in a 190.5 km polar stereographic grid. Data and results provided by R. Brown, Environment Canada.

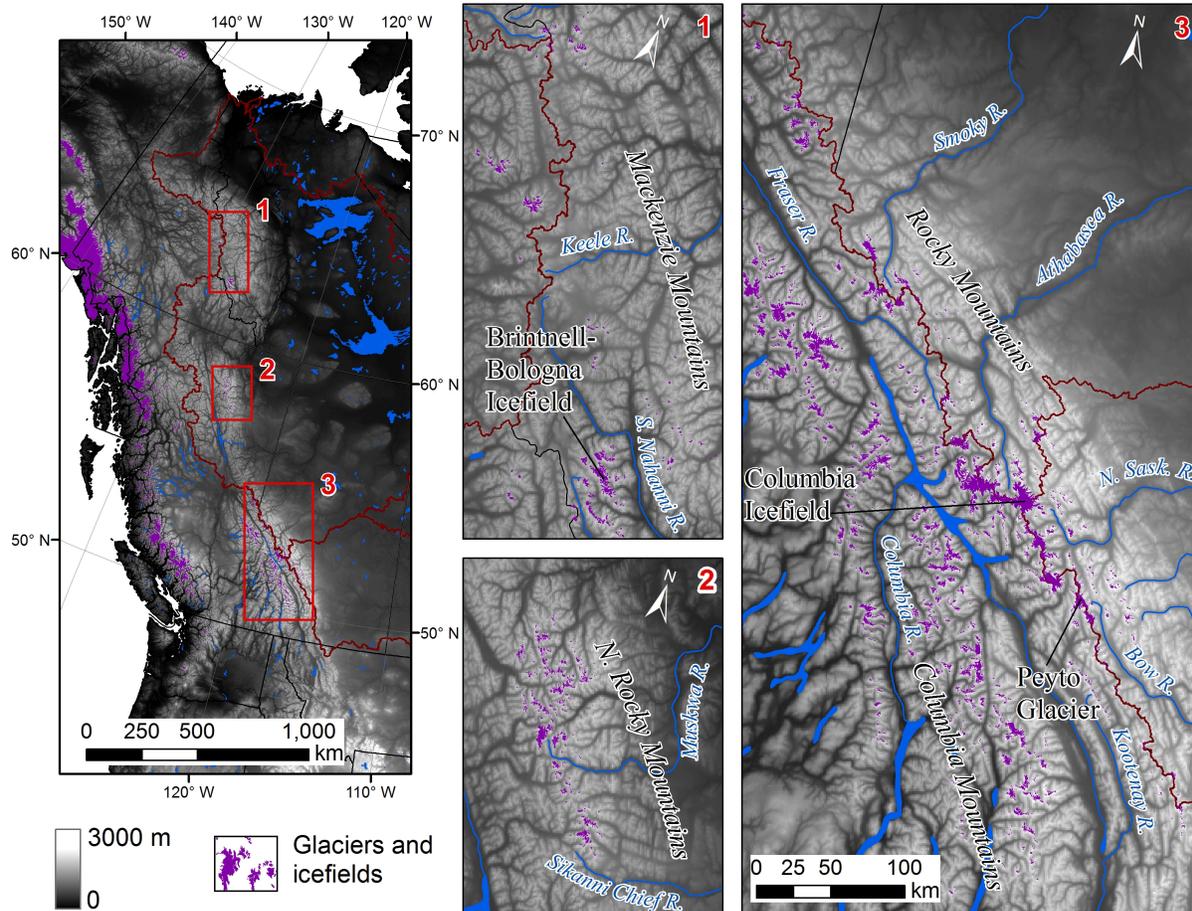


Figure 7. Map of western Canada showing the distribution of glaciers and icefields within the major interior mountain ranges; the locations CCRN glacier observatories are indicated. Glacier extents are from the GLIMS database (www.glims.org/) and the North American Environmental Atlas (www.cec.org/naatlas/).

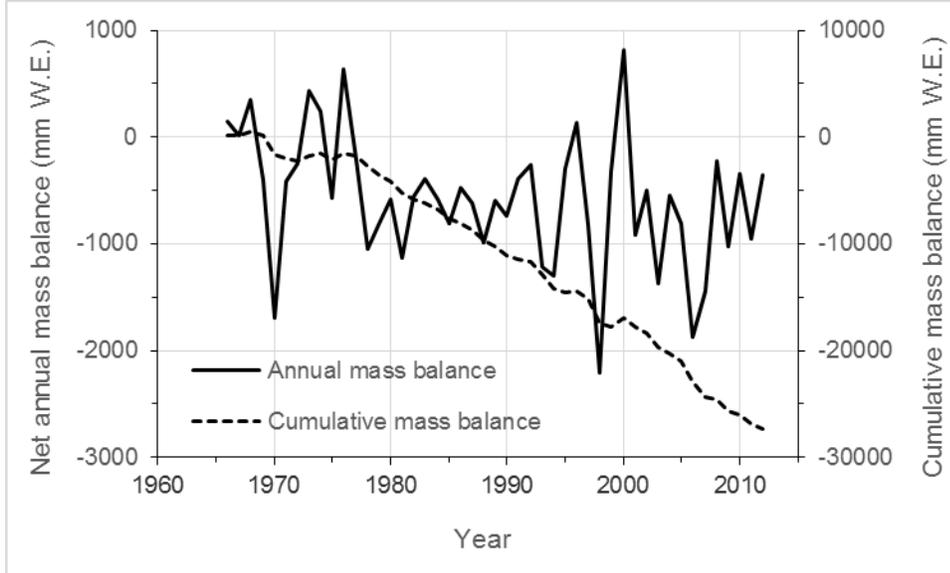


Figure 8. Net annual mass balance series for Peyto Glacier (1966–2012) and cumulative mass balance over the 46-year period. Note that values in 1991/92 are reconstructed from proxy information and values in 2010–12 are preliminary. (Data provided by M. Demuth, Geological Survey of Canada).

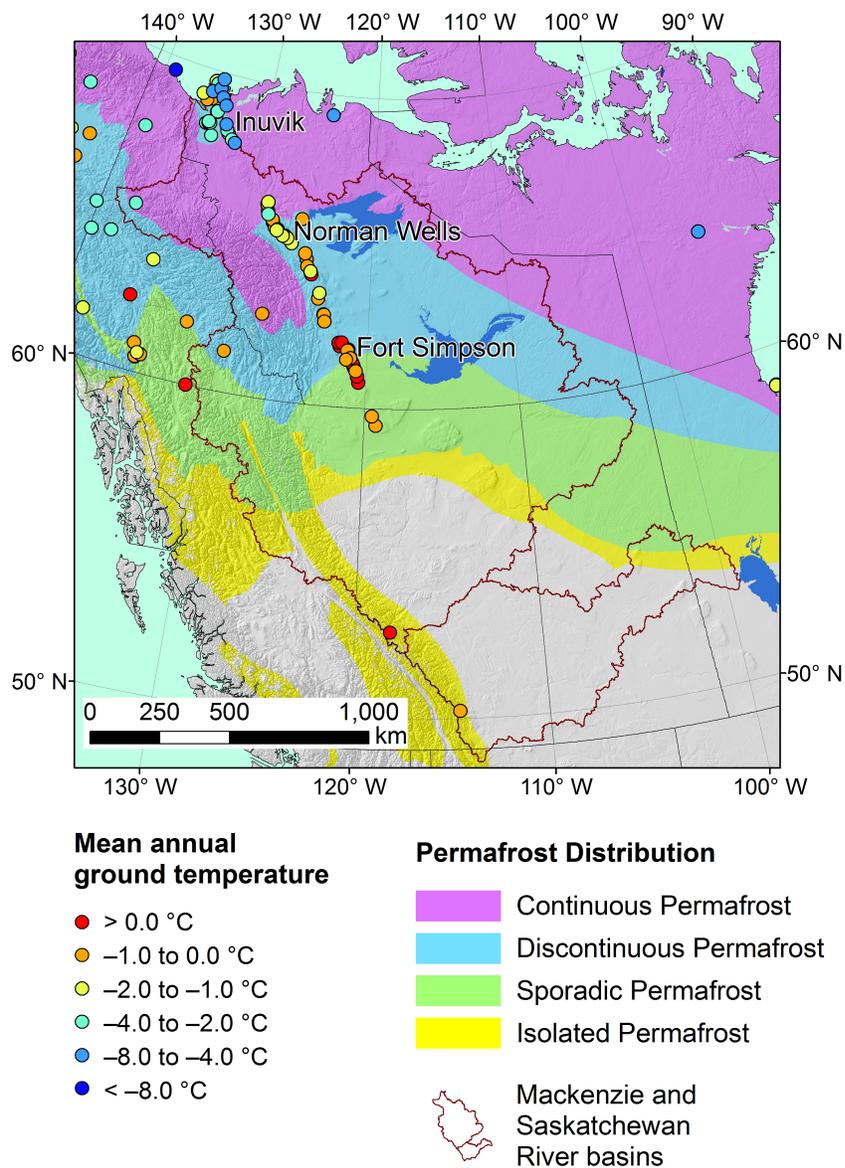


Figure 9. Distribution of permafrost over western Canada (from Brown et al. 2001) and mean annual ground temperature (MAGT) recorded during the IPY (2007–2009) (from IPA, 2010).

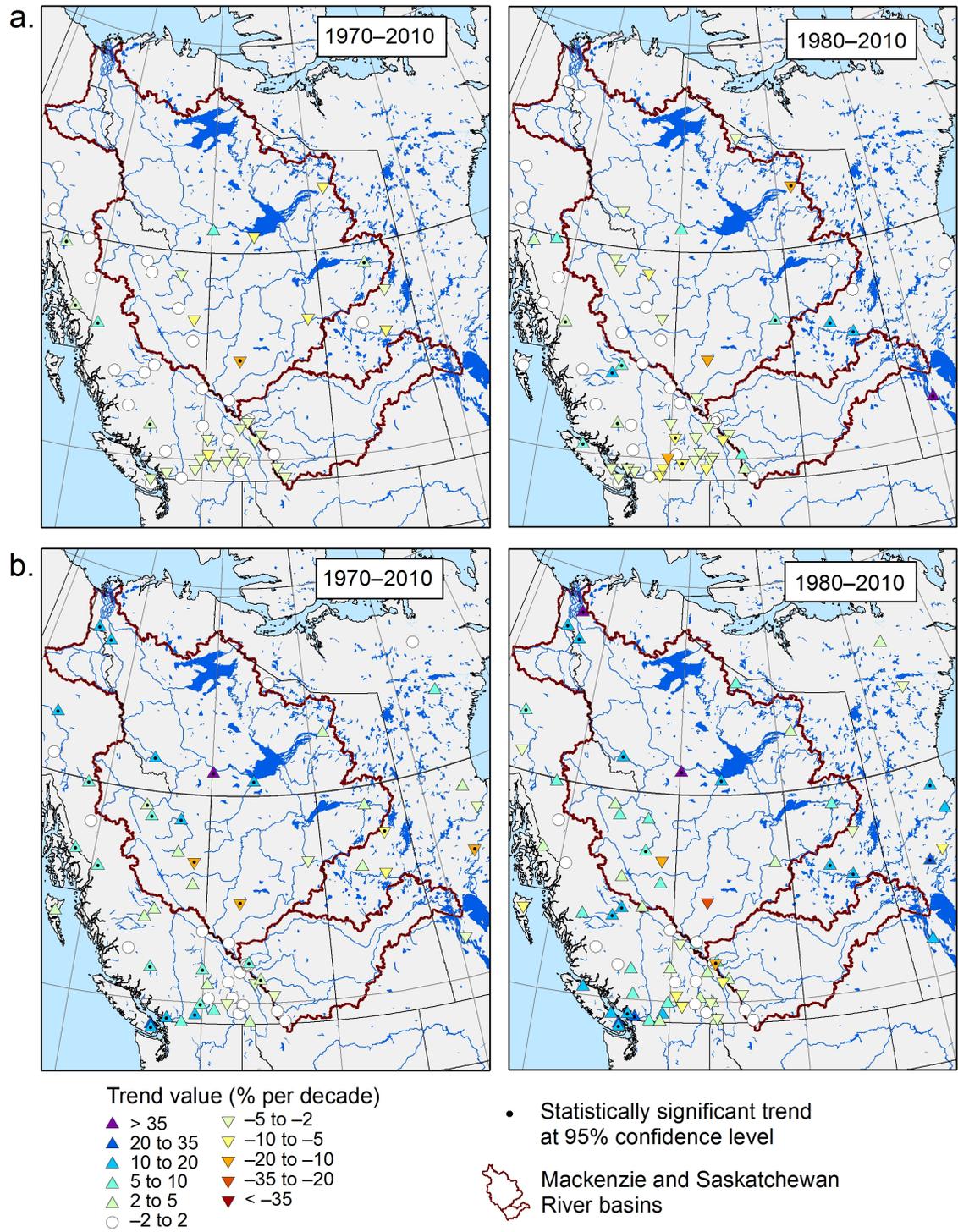


Figure 10. Spatial patterns of trends in a) annual discharge and b) January discharge from RHBN stations over western Canada for the periods 1970–2010 and 1980–2010. Rejected stations are omitted.