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Hydrological resilience of a Canadian Rockies headwaters basin subject to changing climate, extreme weather, and forest management

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Abstract:

Marmot Creek Research Basin in the Canadian Rockies has been the site of intensive streamflow, groundwater, snow accumulation, precipitation, and air temperature observations at multiple elevations. The basin was instrumented in 1962, subjected to forestry experiments in the mid-1970s, and experienced extreme flooding in 2013. Climate change, forest cover change, and recent extreme weather make the basin an ideal laboratory for studying hydrological resilience. Observations show increases in low elevation air temperature, multiple day and spring precipitation, interannual variability of precipitation, and high elevation groundwater levels. Observations also show decreases in peak seasonal snow accumulation and low elevation groundwater levels. Despite these substantial hydrometeorological and groundwater changes, streamflow volume, timing of peak, and magnitude of the peak are not changing. Streamflow volumes are also insensitive to forest cover changes and teleconnections. The June 2013 flood was unprecedented in the period of record, and the basin significantly moderated the hydrological response to the extreme precipitation; the 2013 storm precipitation depth was 65% greater than the next highest storm total over 51 years; however, the 2013 peak streamflow was only 32% greater than the next highest peak flow recorded. The hydrology of Marmot Creek Research Basin displays remarkable resilience to changing climate, extreme weather, and forest cover change. Copyright © 2015 John Wiley & Sons, Ltd.

KEY WORDS climate change; forest hydrology; Canadian Rockies; snow hydrology; resilience; streamflow

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INTRODUCTION

Western Canada and the northwestern USA are highly dependent on water from the Canadian Rockies to meet environmental and socioeconomic needs. Regional population and economic growth will increase demand on water supply from these mountains (Wheater and Gober, 2013). The need to understand the hydrological resilience, i.e., the retention of similar form and function under environmental and human pressures (Creed *et al*., 2011), of these mountain headwater basins is critical in light of increasing water demand and changing climate and forest cover. While there are hydrological sensitivity studies that quantify the change of a system as a function of disturbance to input parameters or meteorological inputs via modelling (e.g. Peterson *et al*., 2012; Botter *et al*., 2013), few observational examples of hydrological resilience exist in the literature.

Manifestations of hydrological resilience often require compensating processes; for example, the extinction of shortwave radiation in a forest canopy compensates for increased longwave emissions from the same canopy in providing energy for snowmelt (Pomeroy *et al*., 2009). Streamflow generation during snowmelt can show resilience to forest disturbance; when clearings are logged, snow accumulation increases dramatically, but snowmelt processes desynchronize with respect to slope and aspect such that basin-scale streamflow generation shows minimal change (Ellis *et al*., 2013). In contrast, non-resilient hydrological systems demonstrate processes that have compounding feedbacks. Observations of a small agricultural basin on the Canadian Prairies showed a very large increase in streamflow, despite no change in annual precipitation volume (Dumanski *et al*., 2015). The streamflow increase was associated with a transition from snowmelt to rainfall dominated runoff generation, clustering in precipitation timing and wetland drainage. The magnitude of hydrological resilience is likely dependent upon the dominant hydrological processes and their relation to climate and biophysical characteristics of the basin. For example, streamflow response to

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changes in precipitation phase partitioning across a selection of basins varies widely (Harder and Pomeroy, 2014). The nonlinear response of hydrological processes to disturbance complicates the prediction of the hydrological resilience of Canadian Rocky Mountain basins, and so the focus of this paper was hydrological resilience to changing climate, teleconnections, flooding, and forest management practices.

Climate change and Canadian Rockies hydrology

It is expected that Canadian Rockies hydrology will be impacted by anthropogenic climate change because of the large temperature increases anticipated in cold regions and the subsequent impact on temperature-sensitive snow processes (Zhang *et al*., 2000; Woo and Pomeroy, 2011). The average warming in western Canada has been 0.5 to 1.5 °C, over 1900 to 1998, with the greatest increases in winter daily minimum temperatures (Zhang *et al*., 2000). With this rise in air temperature, the ratio of rainfall to snowfall is rising in the region (Davis *et al*., 1999; Zhang *et al*., 2000; Shook and Pomeroy, 2012). Increases in the rainfall–snowfall ratio along with more frequent mid-winter melt events have led to a decline in the seasonal snowpack of western North America (Mote *et al*., 2005). Trends in precipitation volume, in contrast to precipitation phase, are contradictory in the Canadian Rockies as some studies show increases (Zhang *et al*., 2000) and others do not (Valeo *et al*., 2007). The observations used in these studies may bias the conclusions as the low elevation location of most long-term climate stations in mountains introduces an elevation bias in the observed trends (Luckman, 1997). Studies in the mountains of southern Idaho, USA, highlight the significance of this bias as they show that hydroclimatic trends vary with elevation. For example, shifts in precipitation phase from snowfall to rainfall and decreases in snowpack depth and duration are greater at lower elevations (Nayak *et al*., 2010).

Streamflow is the integrated response of these hydrometeorological changes at the basin scale. In areas of sparse meteorological monitoring, such as the Canadian Rockies, streamflow is an easily observed indicator that is useful for assessing hydroclimatic change. Many natural annual streamflows in the southern Canadian Rockies appear to have declined in the last century (Rood *et al*., 2005). Some natural hydrographs are exhibiting earlier and lower peak streamflow events, a decline in late summer flows, and an increase in winter flows (Rood *et al*., 2008). However, large-scale main stem streamflow observations do not provide information on the elevation dependency of climate warming in snow-dominated mountain catchments (Tennant *et al*., 2015).

No published studies relate climate change to long-term observations of groundwater levels in the Canadian Rockies. In cold regions, increasing air temperatures and shorter snow-covered periods make more energy available for evapotranspiration and result in drier soils; this may reduce summer recharge of groundwater. Spring recharge, on the other hand, may increase with greater soil infiltrability because of reduced depth and extent of frozen soils (Hayashi and van der Kamp, 2009). Subsurface contributions to mountain rivers and lake inflows are large, ranging from 30% to 100% (Hood *et al*., 2006), and increased groundwater discharge has been identified as the cause of observed increases in river baseflows (Rood *et al*., 2008).

Teleconnections in the Canadian Rockies

Hydroclimatic trends in the Canadian Rockies are most often attributed to climate change but are also modified by cyclical teleconnections such as the Pacific Decadal Oscillation (PDO) and El Nino Southern Oscillation (ENSO). To determine if the observed hydroclimatic trends are due to climate change, rather than cyclical variations associated with teleconnections, it is often argued that teleconnections need to be explicitly quantified (St. Jacques *et al*., 2010). Whitfield *et al*. (2010) and Bonsal *et al*. (2001) provided comprehensive reviews of the PDO and ENSO effects on Western Canadian hydrometeorology. Positive PDO and negative Southern Oscillation Index (SOI; which quantifies ENSO) signals are associated with a low-pressure trough over the North Pacific Ocean and a weak ridge of high pressure over western Canada. This synoptic situation blocks the intrusion of cold arctic air to the continental interior and is associated with decreased precipitation, decreased snow accumulation, decreased streamflow, and increased air temperatures (Whitfield *et al*., 2010; Newton *et al*., 2014). Negative PDO and positive SOI are associated with a ridge of high pressure over the Pacific and a low-pressure trough over the continent. This arrangement leads to the advection of cold arctic air over the continent that correlates to increased precipitation, increased streamflow, and decreased air temperatures (Whitfield *et al*., 2010; Newton *et al*., 2014). The influences of teleconnections are greatest when the quasi-periodic oscillations for PDO (approximately 60-year cycle; St. Jacques *et al*., 2010) and ENSO (approximately 3- to 9-year cycle Bonsal *et al*., 2001) are out of phase (Newton *et al*., 2014). PDO phase changes can only be established 10 to 15 years after a shift because of the variability of the index and long cycle length. The most clearly defined shift in the period of interest (1962 to 2013) occurred in the mid-1970s with a shift from cool to warm phase (Whitfield *et al*., 2010). Recent observations suggest a warm to cool phase shift occurred in the early 2000s (Peterson and Schwing, 2003). Most teleconnection studies assume monotonic relationships between indices and hydrometeorological responses, but nonlinearity is also reported (Fleming and Dahlke, 2014).

Flooding in the Canadian Rockies

Flood events are the most visible manifestation of extreme weather in this region with recent catastrophic floods occurring in 1995, 2005, and 2013. Annual peak flows are constant or declining in much of western Canada (Cunderlik and Ouarda, 2009) but need to be distinguished from extreme flood events that happen infrequently. These extreme events are of great interest as they cause the most damage and occur only under exceptional circumstances. They are also the most difficult to study owing to their infrequent occurrence and typically poor quality data (Whitfield, 2012). This means there is little information to assess whether flood frequency or magnitude is changing. Climate change may increase the probability of extreme weather events that drive these floods (Whitfield, 2012). In the Front Ranges of the Canadian Rockies, extreme floods occur because of large-scale, low-pressure systems blocked by topography (Pennelly *et al*., 2014). Extreme floods can be entirely heavy rainfall driven (e.g. 1995; Alberta Environmental Protection, 1995) or the result of multiple mechanisms acting concurrently, specifically rainfall in addition to snowmelt (e.g. flood of 2013; Pomeroy *et al*., in review). Snowmelt alone is typically unable to generate sufficient runoff rates to cause large floods in this region (Whitfield, 2012). Also important in flood generation in this region, although secondary to atmospheric processes, are antecedent hydrological states such as soil moisture, groundwater level, snowpack, and surface water storage (Pomeroy *et al*., in review).

Forest management and Canadian Rockies hydrology

Forest management practices affect Canadian Rockies hydrology by influencing snow interception, sublimation, evapotranspiration, and melt processes (Pomeroy *et al*., 2012). The removal of forest canopy reduces snow interception and subsequently sublimation of canopy snow (Pomeroy *et al*., 1998). A reduction in sublimation translates directly into an increase in snow accumulation (Pomeroy *et al*., 2012). Snow accumulation in forest clearings is much larger than under the forest (Ellis *et al*., 2010) but is not uniform with maximum accumulations in clearing sizes of two tree heights in diameter (Golding and Swanson, 1978). Increased snow accumulation increases snowmelt volume (Pomeroy and Gray, 1995), which recharges groundwater (Adams *et al*., 1991), and increases both surface and subsurface runoff (Hetherington, 1987). Higher streamflow after logging generally persists until the canopy is re-established (Winkler *et al*., 2010) but varies with differences in runoff generation and evapotranspiration processes (Troendle *et al*., 1987; Granger and Pomeroy, 1997; Pomeroy and Granger, 1997). Suppression of vegetation regrowth after forest removal is required to sustain increased water yields (Macdonald *et al*., 2003). The recovery of the hydrological system depends on the timing of the return of the forest canopy, which itself depends on tree species, planting density, and forest succession and regrowth rates (Winkler *et al*., 2010). At Marmot Creek Research Basin (MCRB), Alberta, the logging of cut blocks increased snow accumulation by 21% and streamflow by 6% (Swanson *et al*., 1986). In contrast, the logging of numerous small clearings increased snow accumulation by 28% and streamflow by 3% to 7% (Swanson and Golding, 1982).

Objectives

The hydrology, hydrometeorology, and land cover of the Canadian Rockies are dynamic. However, studies to date have not described how temperature, precipitation, snow accumulation, groundwater, and streamflow covary, nor how resilient the hydrology of headwater systems is to changes in climate, extreme weather, and land cover. The objective of this paper is therefore to assess comprehensively the hydrological resilience of a mountain basin in relation to changing climate, extreme weather, and forest cover using the uniquely detailed and extensive 50-year MCRB dataset.

SITE AND METHODS

Site

Marmot Creek Research Basin is in the Kananaskis Valley, Alberta, 80 km west of Calgary in the Front Ranges of the Canadian Rockies (Figure 1). The vegetation is typical of the Front Ranges with an elevation gradient from montane forests, subalpine forests, and alpine meadows to sparsely vegetated high alpine talus slopes and exposed bedrock (Swanson *et al*., 1986). Climate is characterized by long cold winters and cool wet summers with monthly average temperatures ranging from 14.1 °C in July to −7.5 °C in January. Average annual precipitation is 638 mm at the valley bottom and around 1100 mm at higher elevations (Storr, 1967).

The snowmelt-dominated streamflow regime of Marmot Creek is a function of MCRB hydrogeology as most streamflow is fed by transient subsurface flow that moderates storm peaks and enhances baseflow (Stevenson, 1971). Water table position is coupled to recharge from rainfall and snowmelt events (Stevenson, 1971).

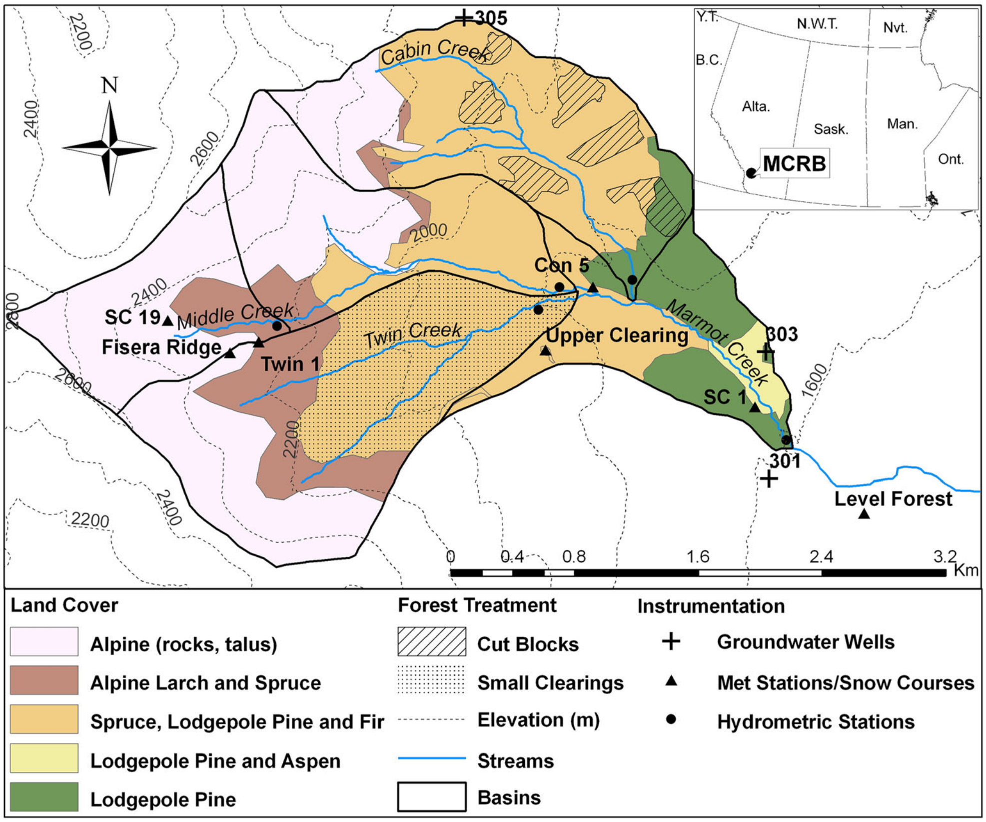

Figure 1. Marmot Creek Research Basin (MCRB) showing selected meteorological and hydrometric stations, groundwater wells, sub-basin boundaries, topographic contours, land cover, forest treatment areas, and location in Canada

Surficial deposits of unconsolidated tills and moraine deposits overlying a bedrock base dominate the hydrogeology, and phreatic divides closely approximate topographic divides. Groundwater generally flows parallel to the slope of topography through talus and scree slopes on valley walls (Toop and de la Cruz, 2002). Flow is generally unconfined, and the water in bedrock has little interaction with that in the overlying high hydraulic conductivity deposits (Stevenson, 1971).

The flood of record in the region occurred on 19–22 June 2013. During this period, portions of southwestern Alberta received over 300 mm of precipitation (Pomeroy *et al*., in review), leading to catastrophic flooding. The majority of precipitation was rainfall, which rapidly filled the available soil storage, and was augmented by snowmelt from rain-on-snow in the upper reaches of the basin. Streamflow observations going back to the late 1800s from Banff and Calgary suggest that this flood, although large, was not unprecedented (Pomeroy *et al*., in review).

Marmot Creek Research Basin has experienced significant anthropogenic impacts from forest manipulation studies conducted by the Canadian Forestry Service during the Marmot Basin Project (1962 to 1986). Different forest harvesting treatments were applied to two sub-basins to discern the impacts of commercial and experimental forest harvesting on water yields, sediment loads, and hydrological regime. In 1974, 23% of the Cabin Creek sub-basin area was logged in six cut blocks ranging from 3 to 13 ha (Beckstead and Veldman, 1985). Between 1977 and 1979, 17% of the Twin Creek sub-basin area was logged in a honeycomb pattern of 2103 15 to 20 m diameter clearings (Beckstead and Veldman, 1985). Middle Creek was unaltered to use as a control. The forestry experiments included detailed hydrometeorological observations from 1962 to 1986 of air temperature, humidity, radiation, wind speed, precipitation, snow accumulation, groundwater level, and discharge. Streamflow measurements occurred at the basin outlet, and at all sub-basin outlets, and meteorological stations and snow courses ranged from low elevation forests to alpine locations. The Canadian Government terminated this programme in 1987 in anticipation of site use for the 1988 Calgary Winter Olympics. Only the basin outlet Water Survey of Canada (WSC) hydrometric station and five Alberta Environment and Sustainable Resource Development groundwater observation wells were operated in MCRB after the Marmot Basin Project. Streamflow observations at the Marmot Creek outlet were year round from 1962 to 1986 and then seasonal (May–October) until June 2013.

The basin was re-instrumented in 2004 by the University of Saskatchewan (U of S) Centre for Hydrology (Pomeroy *et al.*, 2011), hereafter referred to as the U of S Project. The intent of this effort was to improve understanding and modelling of cold region hydrological processes in the Canadian Rockies. Twelve meteorological stations have operated over the course of this project with a subset operating continuously. Site locations range from valley bottom meadows to montane and subalpine forests and to alpine ridges. Hydrometric stations were operated seasonally at each of the same sub-basin stream-gauging sites used during the Marmot Basin Project. The June 2013 flood washed away the sub-basin gauges, and a debris flow infilled the Marmot Creek WSC weir and diverted the channel from this hydrometric station. At the time of writing, no restoration of the WSC weir has occurred, but hydrometric measurements by U of S continue at the WSC gauging site. Snow accumulation measurements occur at monthly intervals during the winter and more frequently during the spring snowmelt season.

Additional weather data comes from a nearby Meteorological Service of Canada station that is currently in a large forest clearing at the University of Calgary Biogeosciences Institute Barrier Lake Field Station (BGSI; 1391 m elevation and 14.5 km northeast of MCRB). The observation period for BGSI is from 1939 to present, with station moves in 1964 and 1976 (Whitfield, 2014). Whitfield (2014) examined the station's fitness for purpose with respect to climate change assessment and concluded that the temperature and precipitation records are suitable for climatological studies at decadal or longer periods and that trend analysis is suitable if the station reinforces trends detected nearby.

Methods

The ability to detect trends in a time series increases with the period of observation and declines with the existence of data gaps. The 1987 to 2004 data gap when research activities were suspended complicates the trend analysis. Because of a lack of appropriate high elevation data in the vicinity, this gap was not infilled to avoid the uncertainty associated with spatial and elevation interpolation. A two-pronged approach deals with the large gap in air temperature, precipitation, maximum snow accumulation, and sub-basin stream flow observations:

1. Trends were quantified from 1962 to 2013 of air temperature, precipitation, groundwater levels, snow, and streamflow. Measurement differences of these variables are small between observation periods. The BGSI air temperature and precipitation records (1939 to 2013) were also analysed for trends.
2. Differences were compared in mean and distribution of variables of interest between the Marmot Basin Project and U of S Project periods. There is uncertainty introduced by the suspension of research activity as sampling locations and measurement techniques changed from 1987 to 2005; for example, in the recent observation period, only automated weather stations are available, while the earlier period included manual observations. Three datasets that span the entire period of interest, including the gap, are Marmot Creek streamflow, MCRB groundwater levels, and BGSI climate data. Change point tests provided context as to hydrometeorological change before, during, and after the gap period.

The Marmot Basin Project sites used in the temperature and precipitation analysis were Confluence 5 (CN5: 1753 m) and Twin 1 (TN1: 2286 m) that correspond spatially to the U of S Project Upper Clearing (UC: large clearing in coniferous forest, 1845 m) and Fisera Ridge (FR: alpine treeline site, 2325 m) sites (Figure 1). Henceforth, the paired low elevation site is termed CN5_UC and the paired upper elevation site is termed TN1_FR. Daily mean (Tmean), maximum (Tmax), and minimum (Tmin) air temperatures were computed for the water year (WY: October to September), winter (October to March), spring (April to June), and summer (July to September) seasons. These seasons correspond to hydrologically distinct periods in the Canadian Rockies. Snow accumulation dominates the winter season, followed by more snow accumulation and ablation, heavy rainfall, and peak streamflow during the spring season, and rainfall and evapotranspiration characterize the summer season (Fang *et al.*, 2013). Without additional information, this analysis assumes the primary difference between the U of S Project and Marmot Basin Project station locations is elevation. Observed lapse rates between the stations correct the U of S Project stations to the Marmot Basin Project station elevations. The nearby BGSI temperature observations were analysed in the same manner as MCRB. Total precipitation for all sites, and rainfall, snowfall, and rain ratios for BGSI, were computed on the same time intervals as the air temperature. WY maximum 1-, 2-, and 3-day precipitation amounts were also calculated. Precipitation data do not account for gauge undercatch because of the lack of appropriate wind data; however, storage gauges had Alter shields in both observation periods in MCRB and a Nipher shield at BGSI. Rain ratio at BGSI was calculated as total rain divided by total precipitation.

Snow accumulation snow water equivalent (SWE) depth was analysed by comparing the WY maximum observed value from transects of snow depth and density measurements paired between the Marmot Basin Project and U of S Project periods. Monthly Marmot Basin Project snow course observations consist of the average of 10 fixed points where snow depth and density were measured (Fisera, 1985). The snow courses surveyed by the U of S Project consist of fixed transects with 13 points measured every 3 m at Level Forest and 30 points measured every 5 m at Fisera Ridge. Density measurements were made every three points.

Table I. Marmot Creek wells

Well	Elevation (m)	Depth (m)	Prod. interval (m)	Aquifer	Lithology
301	1601.4	12.20	6.5–8.5	Rocky mountain	Sandstone
303	1669.1	36.58	34–36.58	Rocky mountain	Sandstone
305	2051.62	11.58	9–11.58	Fernie	Shale

Snow course pairing considered vegetation types, elevations, and observation methodology. The lower forest snow courses paired were the U of S Project Level Forest, a young Lodgepole Pine forest with a 9 m canopy, and the Marmot Basin Project SC#1, an east sloping Lodgepole Pine forest with a 10 m canopy with natural openings (Fisera, 1985). The alpine snow courses paired were the U of S Project Fisera Ridge, a site with north and south facing alpine slope, an alpine ridgetop with extension to the treeline, and the Marmot Basin Project SC#19, a site with north and south facing slopes all situated above the treeline (Fisera, 1985).

Daily groundwater levels from three wells (Table I) were analysed for trends on a WY basis. Parameters calculated for each WY include maximum, minimum, and mean level; differences between maximum and minimum level; standard deviation; maximum, and minimum level date; and the difference between the minimum and maximum level dates. Water years with incomplete records, after linear interpolation of gaps less than 30 days, were removed from the analysis.

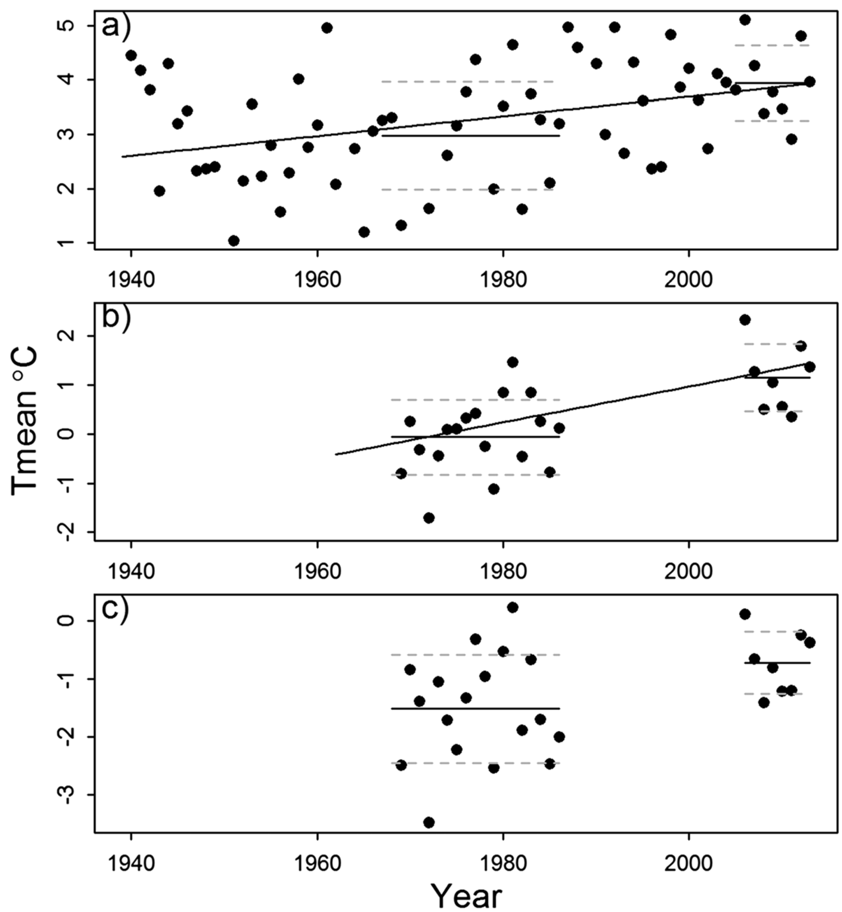

Figure 2. Water year daily mean air temperature at BGSI (a) from 1939 to 2011 and CN5_UC (b) and TN1_FR (c) from 1967 to 2013. BGSI provides temporal and regional contexts for CN5_UC and TN1_FR, which show the difference between the Marmot Creek Project and U of S Project period Tmean observations. The sloping lines in (a) and (b) represent the statistically significant trend. The horizontal black lines are the mean, and the grey dash lines are the standard deviation of their respective periods. BGSI, Biogeosciences Institute; CN5, Confluence 5; UC, Upper Clearing; TN1, Twin 1; FR, Fisera Ridge

Streamflow gauging used the application of rating curves to continuous stage measurements. Streamflow during the Marmot Basin Project period was estimated year round using stage records from flow through V-notch weirs on Marmot, Middle and Twin Creeks, and an H-flume on Cabin Creek. After 2006, sub-basin streamflow was gauged using hourly pressure transducer measured stage and periodic acoustic Doppler velocity measurements in natural channel reaches. The U of S Project instrumentation was digital, and rating curves came from 15+ site visits per year. However, because of the natural reaches, these recent estimates are inherently less reliable than from the weirs of the earlier Marmot Basin Project. Of the four sub-basins (Twin, Cabin, Middle, and Alpine), two had forest cover manipulations (Twin and Cabin) and two are considered controls (Middle and Alpine) (Figure 1). An estimate of peak daily flow during the 2013 flood came from the only streamflow data in the basin to survive the event, a stage recording of the alpine portion of the Middle Creek. Regression of streamflow between Middle Creek and Marmot Creek estimated the streamflow at the Marmot Creek outlet. The assessment of streamflow change considered the following:

1. trends in May–October monthly and seasonal average streamflow for each sub-basin,
2. peak flow magnitude and date, and
3. timing of the 25th, 50th, and 75th percentiles of seasonal cumulative flow over May–October.

To identify the impact of the forestry manipulation, the analysis considered the following:

1. WY, monthly (May to October), peak, and low flow (10th percentile streamflow) runoff ratios in the experimental (Twin or Cabin Creek) and control (Middle Creek) basins.
2. Differences in timing of the peak flow, and the 25th, 50th, and 75th percentiles of cumulative WY flow.

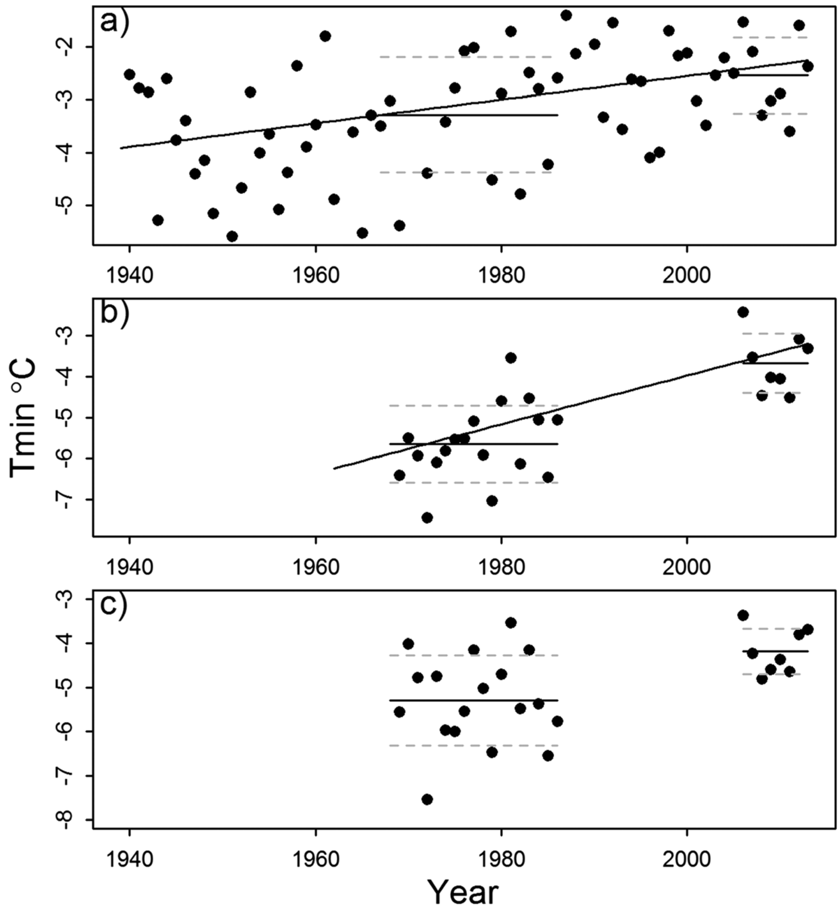

Figure 3. Water year daily minimum air temperature at BGSI (a) from 1939 to 2011 and CN5_UC (b) and TN1_FR (c) from 1967 to 2013. BGSI provides temporal and regional contexts for CN5_UC and TN1_FR, which show the difference between the Marmot Creek Project and U of S Project period Tmin observations. The sloping lines in (a) and (b) represent the statistically significant trend. The horizontal black lines are the mean, and the grey dash lines are the standard deviation of their respective periods. BGSI, Biogeosciences Institute; CN5, Confluence 5; UC, Upper Clearing; TN1, Twin 1; FR, Fisera Ridge

Studies linking PDO and ENSO teleconnections to the hydroclimatology of the region (Bonsal *et al*., 2001; St. Jacques *et al*., 2010; Fleming and Sauchyn, 2013) motivated an analysis of the relationship between these indices and the detailed MCRB hydrometeorological observations. The analysis follows the generalized least square regression (GLS) approach of St. Jacques *et al*. (2010).

Statistical tests. The Mann–Kendall (MK) test evaluated the presence and significance of trends, while the Theil–Sen estimator quantified the magnitude of change. The MK test is a non-parametric test that assesses if a time series exhibits a monotonic trend or not; for further information, see Yue and Pilon (2004) and Yue *et al*. (2002a). Autocorrelation, common in hydroclimatic datasets, increases the probability of the MK test detecting significant trends. To reduce the influence of autocorrelation, time series are often pre-whitened (Yue *et al*. 2002b). As large uncertainties exist in the calculation of autocorrelation in the presence of gaps, data with and without pre-whitening were assessed for trends with the MK test. Few differences were detected; thus, pre-whitened results are presented to be consistent with much of the literature on this topic. The Theil–Sen estimator is a non-parametric statistic that estimates the median slope of a dataset based on pairs of sample points (Sen, 1968). In addition, the non-parametric Kolmogorov–Smirnov (KS) and Mann–Whitney *U* (MW) tests assessed the significance of the difference between distributions for the Marmot Basin Project (1965 to 1986) and U of S Project (2005 to 2013) periods. To determine if there were specific points in time that correspond to shifts in the hydrometeorology, prior, during, or post data gap, a change point detection analysis was implemented. A maximum likelihood-ratio approach estimates the point of change in the mean value of a distribution and its probability (Hinkley, 1970). This method assumes a normal distribution and identifies at most one change (Killick and Eckley, 2014). Only significant changes are reported in the results.

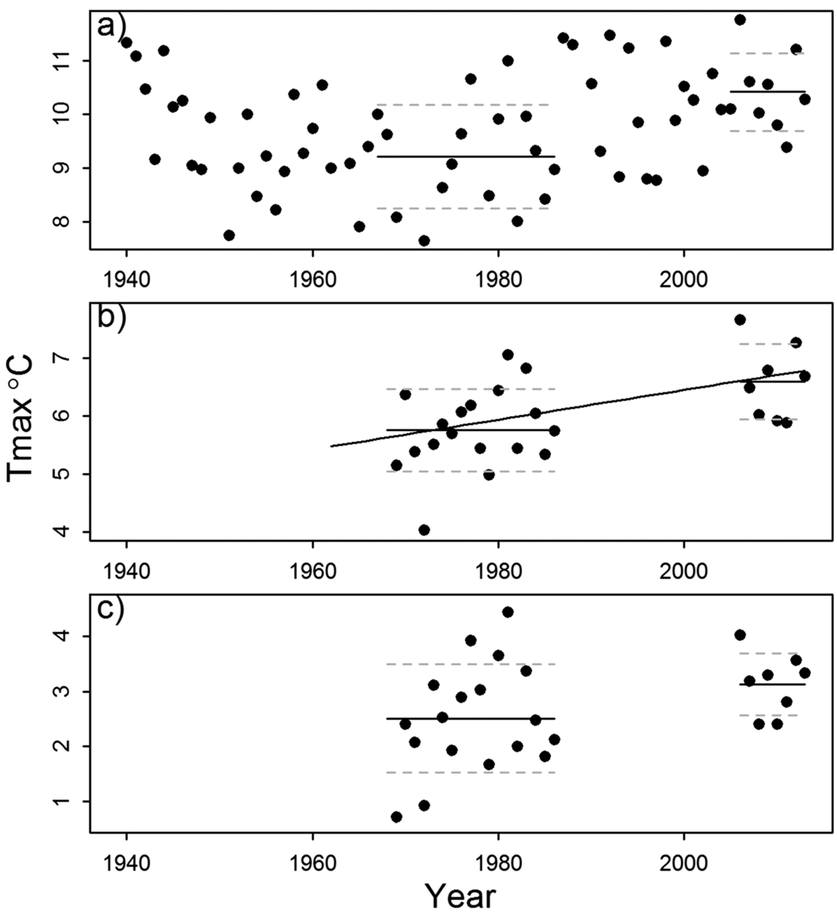

Figure 4. Water year daily maximum air temperature at BGSI (a) from 1939 to 2011 and CN5_UC (b) and TN1_FR (c) from 1967 to 2013. BGSI provides temporal and regional contexts for CN5_UC and TN1_FR, which show the difference between the Marmot Creek Project and U of S Project period Tmax observations. The sloping line in (b) represents the statistically significant trend. The horizontal black lines are the mean, and the grey dash lines are the standard deviation of their respective periods. BGSI, Biogeosciences Institute; CN5, Confluence 5; UC, Upper Clearing; TN1, Twin 1; FR, Fisera Ridge

The teleconnection analysis used a GLS model that included terms for the trend of the variable of interest, PDO, ENSO, and the residual quantified by an autoregressive moving average process (ARMA). PDO data come from Mantua (2015), while SOI data come from the Australian Bureau of Meteorology (2015). Mean winter averages (November to March) for PDO and SOI were related to mean value of the variable of interest in the subsequent year (Bonsal *et al.*, 2001; St. Jacques *et al.*, 2010). To develop statistical models of the variables of interests, the smoothing and order of the ARMA error term were perturbed. Comparison of the resulting models with the Akaike Information Criterion statistic (Sakamoto *et al.*, 1986) identified best-fit. Statistical significance of the terms was determined with the Neyman–Pearson statistic (Zheng *et al.*, 1997), a likelihood-ratio test that compares the models with and without the forcing term of interest included.

To highlight some characteristics of floods in MCRB, annual peak daily streamflows exceeding the 90% cumulative probability of the fit generalized extreme value distribution ($\approx 1.8\ m^3/s$) are classified as floods. Estimates of precipitation depths associated with floods came from MCRB gauges or from nearby weather stations when there were no MCRB precipitation data.

The statistical significance criterion for this analysis is $p = 0.05$. All statistical tests, including GLS (nlme package, Pinheiro *et al.*, 2012), MK (Kendall package, McLeod, 2011), autocorrelation pre-whitening (ZYP package, Bronaugh and Werner, 2013), change point

Table II. Air temperature statistics

Station	Interval	Trend (°C/year)[a]	1967–1986 Mean (°C)	2005–2013 Mean (°C)	Difference (°C)	Change point (year)[b]
Tmin						
BGSI	WY	0.022	−2.95	−2.50	0.45	1974
BGSI	Winter	0.028	−8.51	−7.70	0.81	
BGSI	Summer	0.011	4.90	5.26	0.36	1969
CN5_UC	WY	0.060	−5.65	−3.67	1.98[c]	
CN5_UC	Winter	0.077	−11.89	−9.54	2.35[c]	
CN5_UC	Summer	0.076	2.81	5.41	2.60[c]	
TN1_FR	WY		−5.30	−4.19	1.11[c]	Gap
TN1_FR	Winter		−10.88	−9.95	0.93[e]	1979
TN1_FR	Summer		2.99	5.04	2.05[c]	
Tmean						
BGSI	WY	0.018	3.23	3.81	0.58[c]	1986
BGSI	Winter	0.019	−3.11	−2.13	0.98[c]	1985
BGSI	Summer	0.013	11.79	13.40	1.61[d]	1993
CN5_UC	WY	0.037	−0.06	1.16	1.22[c]	Gap
CN5_UC	Winter	0.049	−6.89	−5.34	1.55[c]	
CN5_UC	Summer		9.16	11.12	1.96[c]	
TN1_FR	WY		−1.52	−0.73	0.79[d]	1975
TN1_FR	Winter		−7.52	−6.74	0.78[e]	1976
TN1_FR	Summer		7.24	9.10	1.86[c]	
Tmax						
BGSI	WY		9.20	10.28	1.08[a]	1986
BGSI	Winter		2.20	3.33	1.13[d]	1946
BGSI	Summer		19.16	21.39	2.23[d]	1993
CN5_UC	WY	0.026	5.75	6.59	0.84[c]	1979
CN5_UC	Winter	0.036	−1.68	−0.65	1.03[c]	1979
CN5_UC	Summer		15.89	17.85	1.96[c]	
TN1_FR	Winter		−4.03	−3.23	0.80[e]	
TN1_FR	Summer		11.97	13.55	1.58[c]	Gap

BGSI, Biogeosciences Institute; MCRB, Marmot Creek Research Basin; CN5, Confluence 5; UC, Upper Clearing; TN1, Twin 1; FR, Fisera Ridge; MK, Mann–Kendall; KS; Kolmogorov–Smirnov.
BGSI trends are for the period 1939–2013, while MCRB trends are for 1967–2013. WY refers to a 'water year' starting 1 October.
[a] MK significant at $p \leq 0.05$.
[b] Significant at $p \leq 0.05$, gap refers to a change point detected in the gap period 1986–2005.
[c] KS and MW significant at $p \leq 0.05$.
[d] MW significant at $p \leq 0.05$.
[e] KS significant at $p \leq 0.05$.

Table III. Precipitation statistics

Station	Interval	Variable	Trend (mm/year)[a]	1967–1986 Mean (mm)	2005–2013 Mean (mm)	Difference
BGSI	WY	Precip		624.1	732.3	108.2[c]
BGSI	Spring	Precip		218.3	359.5	141.2[b]
BGSI	WY	Rain		353.2	512.9	159.7[c]
BGSI	Spring	Rain		129.3	250.8	121.5[b]
BGSI	Spring	Snow	0.5	87.7	75.1	−12.6
BGSI	WY	Portion rain[e]		0.55	0.66	0.11[b]
BGSI	WY	3-day acc.	0.3	66.4	117.1	50.7[b]
CN5_UC	Spring	Precip	1.3	221.2	290.6	69.4
TN1_FR	Spring	Precip		280.8	399.1	118.3[d]

BGSI, Biogeosciences Institute; WY, water year; CN5, Confluence 5; UC, Upper Clearing; TN1, Twin 1; FR, Fisera Ridge; MK, Mann–Kendall; KS; Kolmogorov–Smirnov.
BGSI trends are for the period 1939–2011, while MCRB trends are for 1967–2013.
[a] MK significant at $p \leq 0.05$.
[b] KS and MW significant at $p \leq 0.05$.
[c] MW significant at $p \leq 0.05$.
[d] KS significant at $p \leq 0.05$.
[e] Dimensionless as variable is a ratio of rain/total precipitation.

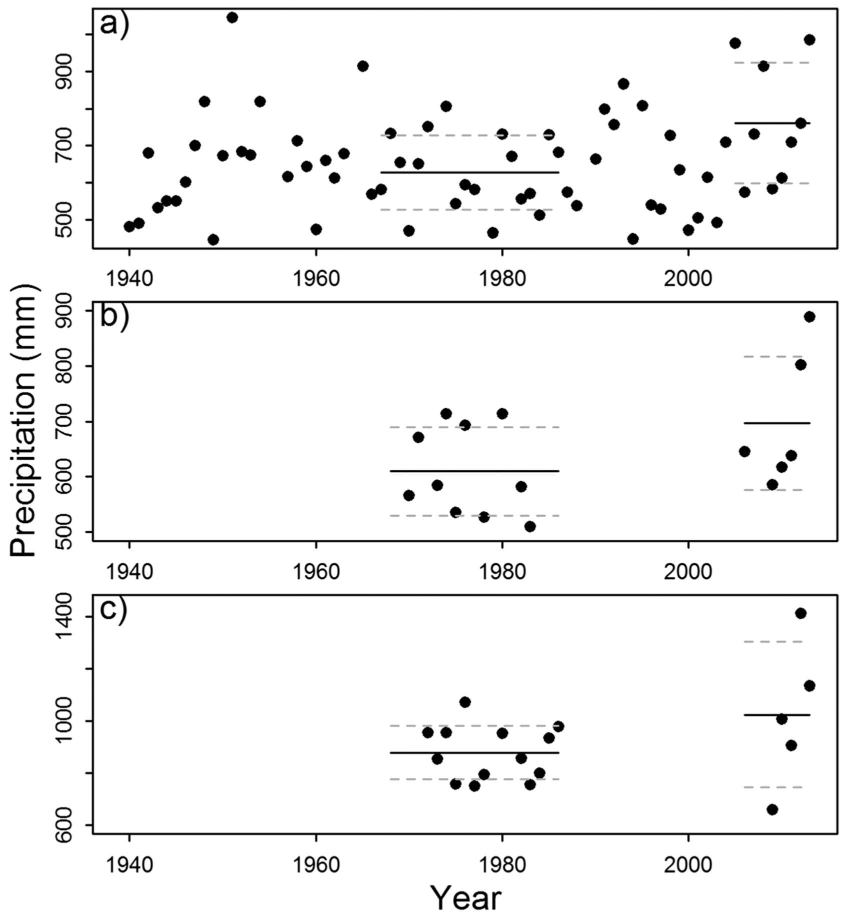

Figure 5. Water year precipitation amount (mm) at BGSI (a) from 1939 to 2011 and CN5_UC (b) and TN1_FR (c) from 1967 to 2013. BGSI provides temporal and regional contexts for CN5_UC and TN1_FR, which show the difference between the Marmot Creek Project and U of S Project period precipitation observations. The horizontal black lines are the mean, and the grey dash lines are the standard deviation of their respective periods. BGSI, Biogeosciences Institute; CN5, Confluence 5; UC, Upper Clearing; TN1, Twin 1; FR, Fisera Ridge

(Changepoint package, Killick and Eckley, 2014), and Theil–Sen (WRS package, Wilcox and Schönbrodt, 2013), were computed using the open-source statistical program R (Ihaka and Gentleman, 2011).

RESULTS AND DISCUSSION

Observed trends with respect to elevation

Low elevation observations range from the base of the basin from 1600 to 1845 m, while upper elevation observations range from 2051 to 2832 m and include the treeline at 2300 m. This discussion of hydrometeorological change considers upper elevations to be greater than 1900 m and lower elevations to be less than 1900 m.

Temperature. The trends in MCRB air temperature from 1967 to 2013 (Figures 2–4 and Table II) vary with elevation. Increases in Tmin occur more often than increases in Tmean or Tmax. Six Tmin WY or seasonal intervals showed warming trends *versus* five and two trends for Tmean and Tmax, respectively (out of a possible of 12 intervals). The magnitude of warming was also the greatest for Tmin. For example, from 1967 to 2013 at CN5_UC, winter Tmin increased by 3.6 °C *versus* 2.3 °C for Tmean and 1.7 °C for Tmax. This pattern is consistent across all other intervals and locations. Spring air temperatures did not show trends at any location or interval. Greater warming occurred at lower elevations (CN5_UC) than at upper elevations (TN1_FR).

The trends reported from the valley bottom BGSI site span a longer period (1939–2013) and have few data gaps. Like MCRB, the greatest warming at BGSI occurs in minimum temperatures, with WY Tmin and Tmean having increased 1.7 and 1.4 °C, respectively, while Tmax did not change. This is consistent with the findings of Whitfield (2014) at BGSI. The warming was greater in winter (Tmin 2.1 °C) than in summer (Tmin 0.8 °C), and there were no trends in spring. The MCRB trends are consistently greater in magnitude than those for BGSI. The WY Tmin increase

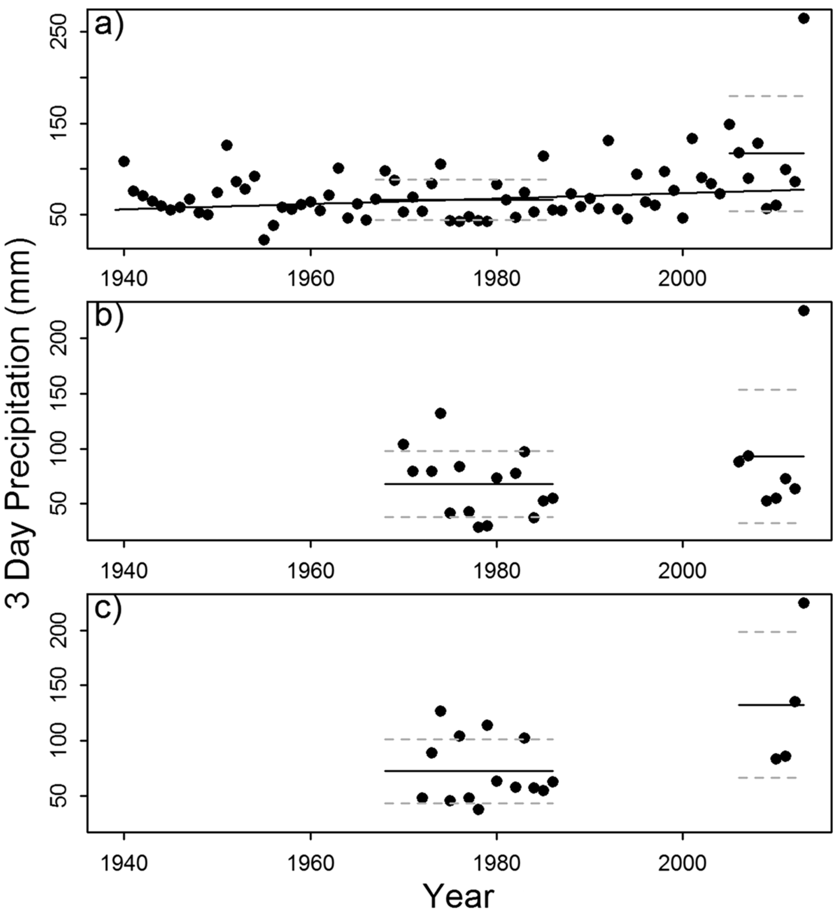

Figure 6. Maximum 3-day precipitation amounts (mm) at BGSI (a) from 1939 to 2011 and CN5_UC (b) and TN1_FR (c) from 1967 to 2013. BGSI provides temporal and regional contexts for CN5_UC and TN1_FR, which show the difference between the Marmot Creek Project and U of S Project period precipitation observations. The sloping line in (a) represents the statistically significant trend. The horizontal black lines are the mean, and the grey dash lines are the standard deviation of their respective periods. BGSI, Biogeosciences Institute; CN5, Confluence 5; UC, Upper Clearing; TN1, Twin 1; FR, Fisera Ridge

of 2.7 °C at CN5_UC over 47 years is much greater than the increase of 1.7 °C at BGSI over 75 years. This is a consequence of a shorter time series not effectively constraining the Theil–Sen slope or is due to the enhanced exposure of higher elevations to the atmosphere at MCRB. Change points were identified from the 1970s to 1990s with little consistency between the different temperature variables, time intervals, or sites. Change points did not correspond to the station moves in 1964 and 1976 reported in Whitfield (2014), suggesting that concerns with possible overestimation of Tmin trends at BGSI due to station moves were not realized.

Precipitation. The only precipitation trends identified were increasing spring precipitation at the low elevation CN5_UC (1.3 mm/year) and declining spring snowfall at the similarly low elevation BGSI (0.5 mm/year) (Table III). Large differences between the Marmot Basin Project and the U of S Project periods were due to increases in spring precipitation at all MCRB elevations and BGSI. Despite the lack of trends, there is an increase in annual precipitation and interannual variability in precipitation between the two observation periods at MCRB (Figure 5). Trend analysis of the maximum multi-day precipitation at all sites identified the 3-day maximum precipitation to be increasing at BGSI. A substantial increase in 3-day maximum precipitation between the observation periods in MCRB only occurred at the upper elevation site (Figure 6). These increases in multiple day precipitation are consistent with those found by Shook and Pomeroy (2012) across the Prairie Provinces of Canada. Gaps in the dataset at MCRB often caused the statistical analysis to not identify trends even when large differences exist between the early and later observation periods. No change points were detected in any precipitation record.

Snowpack. The elevation dependency of the temperature trends strongly reflects observations of peak SWE. The upper elevation (alpine) snow course (Figure 7a) showed no change in peak SWE between two statistically similar periods (KS and MW tests); 1963 to 1986 mean peak SWE was 251 mm and 2005 to 2013 mean peak SWE was 229 mm. The 55% decline in the lower elevation peak SWE from 1963 to 1986 is consistent with recent low peak SWE observations (Figure 7b); mean peak SWE is 62 and 52 mm for 1981 to 1986 and 2006 to 2013, respectively. The implication is that most of the 82% decline in peak SWE over 1963 to 2013 at this upper elevation happened prior to 1986. The decline in peak SWE at the lower elevations between 1963 and 1986 is consistent with increases in winter air temperatures at lower elevations that could result in increased rainfall *versus* snowfall and mid-winter ablation events. No change points were detected in precipitation.

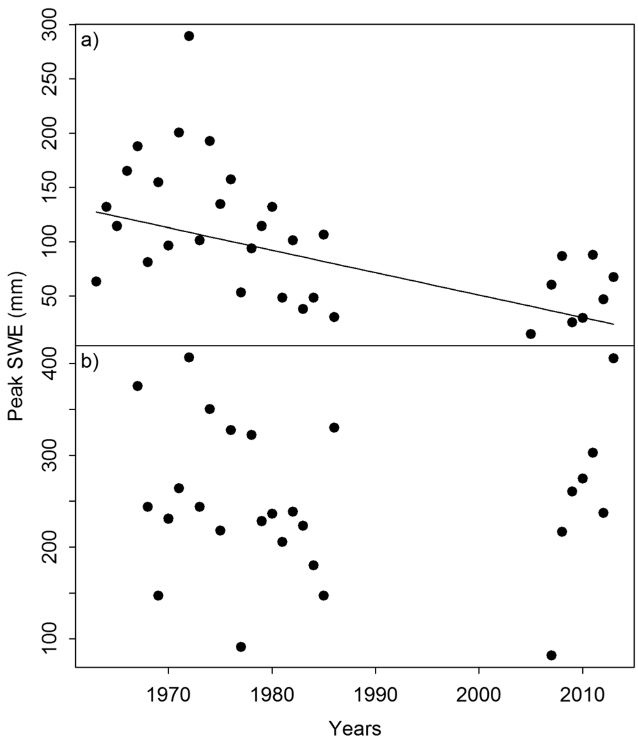

Figure 7. Marmot Creek peak snow accumulation observed snow water equivalent (SWE) for the lower forest (a) and alpine (b) snow courses. The sloping black line in (a) is the observed significant trend. For the respective periods, the solid horizontal black lines are the mean values, and the grey dashed lines are the standard deviations. The Kolmogorov–Smirnov and Mann–Whitney tests confirm no significant difference in distribution between Marmot Creek Project and U of S Project periods for the alpine site (b)

Groundwater. Trends in water table observations vary with elevation, location with respect to nearby creeks, and depth (Table IV and Figure 8). The shallow well at low elevation near to Marmot Creek shows no change (Well 301), whereas a deeper well at low elevation further from Marmot Creek shows a decline (Well 303), and a shallow well at high elevation away from any creek shows an increase (Well 305). It is important to note that the screen elevation of Well 303 is at a higher elevation than Well 301. The most likely explanation of these trends is that (1) Well 301 shows the influence of hyporheic exchange with

Table IV. Groundwater statistics

Well	Variable	Trend (m/year)[a]	Change over period (m)	Change point (year)[a]
303	Min	−0.043	−2.09	
303	Mean	−0.012	−0.57	
303	SD	0.021	1.05	1989
305	Max	0.037	1.82	1992
305	Min	0.038	1.87	1973
305	Mean	0.018	0.9	1971

[a] Significant at $p < 0.05$.

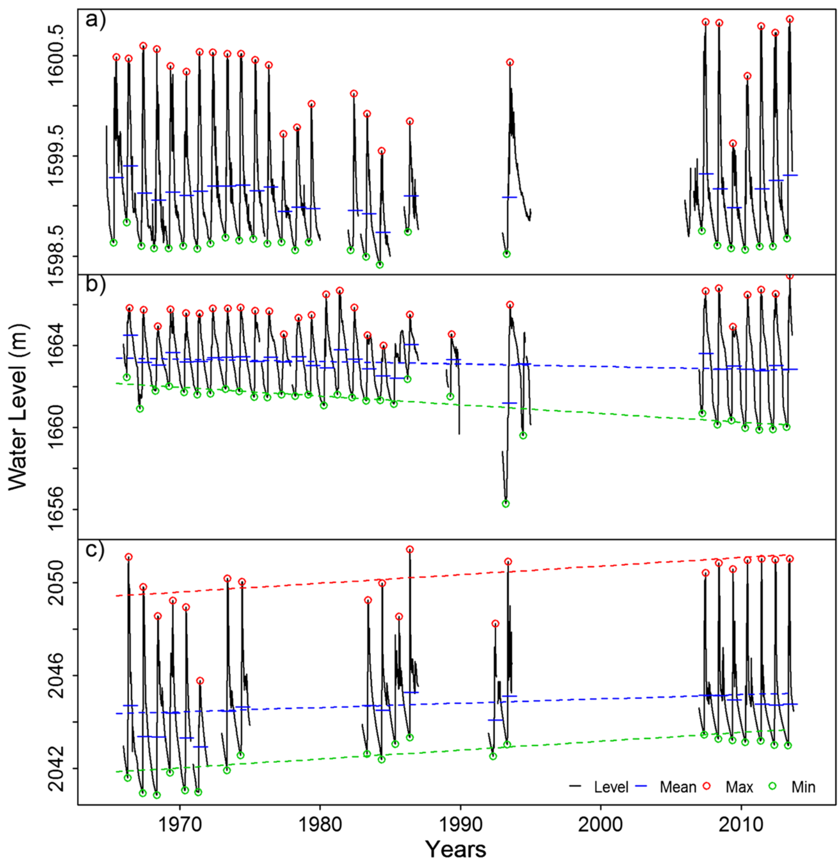

Figure 8. Water table levels for (a) Well 301 (1601 m), (b) Well 303 (1669 m), and (c) Well 305 (2052 m) with water year peak levels (red circles), minimum levels (green circles), and mean levels (blue lines). The statistically significant trends are identified with dashed lines for mean levels (blue), minimum levels (green), and maximum levels (red)

Marmot Creek, (2) Well 303 reflects a decline in low elevation snow accumulation and its effect on recharge, and (3) Well 305 displays the cumulative effect of increasing high elevation multiple day precipitation events and non-declining snow accumulation increasing recharge. Because of the shallow depth (12 m) of Well 305 and the close proximity to a nearby early 1970s clear-cut, the signal may reflect changes in land use; change points are detected in 1973 and 1971 for minimum and mean groundwater levels, respectively.

Streamflow. There were no trends or change points identified for Marmot Creek streamflow for monthly or seasonal volumes, peak magnitude, timing of peak, and various flow volume percentiles (25^{th}, 50^{th}, and 75^{th}). An analysis of the statistical significance and magnitude of trends as they change over time highlights, and provides an instructive example, of the sensitivity of the MK test to the selected time period (Figure 9b and c and Table V). For instance, trends were found for the Marmot Basin Project Period for 1963 to 1986 (−27%) and also 1962 to 2010 (−24%) (Figure 9a). If analysis had been limited to periods that started in 1963 and ended between 1985 and 1990, or between 2002 and 2011, the reported result would be a statistically significant decline in streamflow. However, when considering the full time series, there are no statistically significant trends. The visualization of the changing statistical significance of the trends highlights two important points. First, studies that present findings of trend analysis only represent conditions up to the time of the last data point. Recalculation of climatic and hydrological change in subsequent years can yield changing conclusions as climate and land cover are constantly changing. Carefully quantified statistical significances are lost via the inclusion of two particularly wet or dry years. Second, to assess environmental change and hydrological resilience, research basins need to operate continuously over time.

Only the Cabin Creek sub-basin had a significant trend in streamflow with a decline in peak flow. This decline was not matched by any other trends in other sub-basins, suggesting that the resilience in the basin-scale streamflow generation processes is consistent across scales within the basin. As the various sub-basins represent distinct elevation zones, this also suggests that there is a multi-elevation resiliency in streamflow

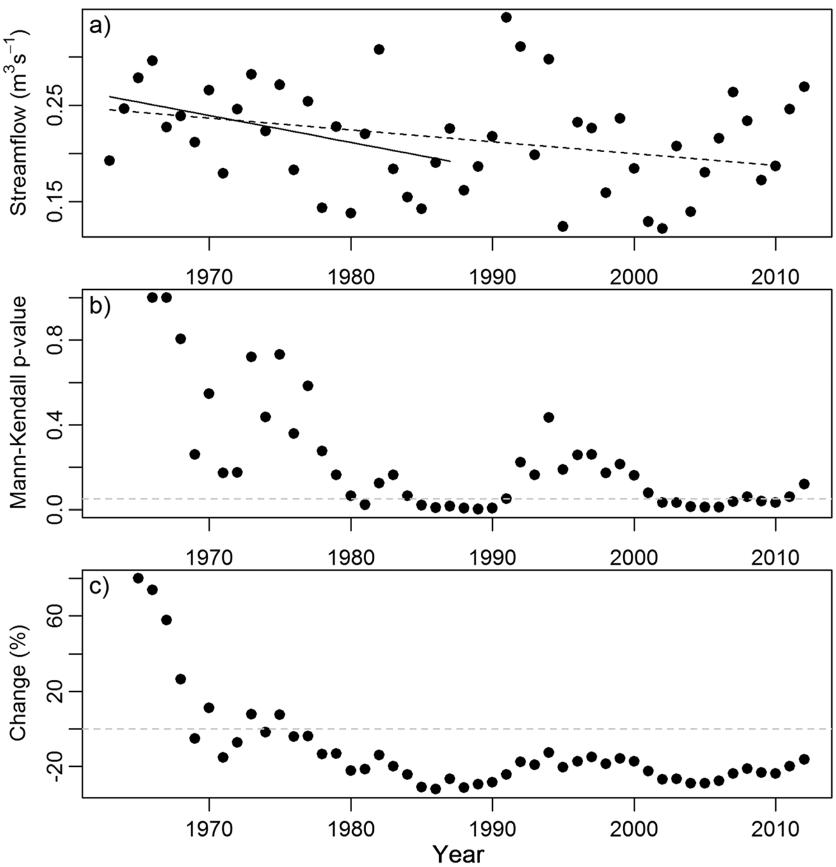

Figure 9. Variation in (a) Marmot May–October average streamflow (1962–2012), (b) statistical significance (Mann–Kendall *p* value), and (c) trend magnitude as calculated on an annual basis from 1965 to 2012. Dashed line in (a) represents trend for 1962–2010, and solid line represents trend for 1963–1986. Dashed line in (b) represents the 0.05 statistical significance thresholds, and dashed line in (c) represents 0% (no trend). A minimum of three data point is required to calculate a Mann–Kendall *p* value so the left most data points in (b) and (c) represent the period from 1963 to 1965, the next point represents 1963–1966, and so on

Table V. Streamflow trend analysis

Basin	Variable	Period	MK *p* value	Trend (m^3/s year)[a]	Change over period (%)
Marmot	Streamflow[a]	1962–2012	0.12	−0.0008	−16
Marmot	Streamflow[a]	1962–2010	0.033	−0.0012	−24
Marmot	Streamflow[a]	1962–1986	0.013	−0.0027	−27
Cabin	Peak streamflow	1963–1986	0.041	−0.0085	−72

MK, Mann–Kendall.
[a] May–October mean streamflow.

generation under changing climate. This result differs from Tennant *et al*. (2015) who, on a larger scale, found an elevation dependency of climate warming impacts in snow-dominated mountain basins.

Summary. Trend analysis of MCRB shows that snow accumulation is declining at low elevations but not at high elevations. This contrasts with many studies of western North American climate change impacts on catchment hydrology that suggest snow accumulation declines at all elevations (Stewart *et al*., 2004; Mote *et al*., 2005; Rood *et al*., 2005) and demonstrates the importance of examining accumulation at a range of elevations. For instance, Nayak *et al*. (2010) considered elevation effects in a trend analysis of the Reynolds Creek, Idaho mountain catchment, and their findings of a greater decline in snowpack at lower and mid elevations were consistent with MCRB. However, a corresponding decline in high elevation streamflow at Reynolds Creek (Nayak *et al*., 2010) did not manifest itself at MCRB, suggesting greater hydrological resiliency to high elevation climate change at MCRB. It is notable that the inclusion of two recent, wet years in the streamflow analysis led to a different conclusion than those reached in other studies of this region that ended in drier periods, e.g., Rood *et al*. (2005), St. Jacques *et al*. (2010), and Stewart *et al*. (2004). Overall, the results indicate that the hydrological system of MCRB is resilient to the observed changes in climate over the past half-century.

Flood

There is no temporal trend in annual peak daily streamflow magnitude or timing for Marmot Creek over the period of record (Figure 10). All floods occurred in June, a period associated with large amounts of precipitation concomitant with the active snowmelt period. All but one flood (1974) occurred in association with large rainfall events. The snowmelt-driven 1974 flood was local to MCRB and did not result in large-scale downstream flooding. Both the precipitation and streamflow records show that 2013 had both the largest daily discharge magnitude on record and the latest peak flow timing (Table VI). The 2013 event peak streamflow was 32% greater than the previous streamflow record (1995), while the 2013 event precipitation measured at BGSI was 65% greater than the next highest event (2005). This demonstrates hydrological resilience due to dampening of the impact of precipitation extremes on streamflow in MCRB due to the storage and moderated release of precipitation from the subsurface. Analysis of the timing and magnitude of these floods events, as well as annual peaks, shows no trends with time. The shifts in flood generation mechanisms in the region described by Pomeroy *et al*. (in review) occurred in the early 20th century, well before this observation period.

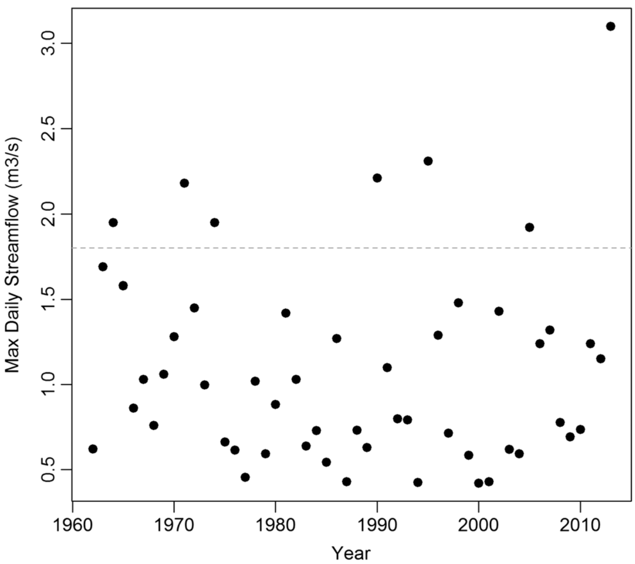

Figure 10. Marmot Creek Research Basin peak daily discharge (m^3/s). The dashed line is the threshold to identify flood events and represents the 90% cumulative probability of the fitted generalized extreme value distribution (1.8 m^3/s)

Forestry manipulation

The assumption of Swanson *et al*. (1986) that the statistical significance of a streamflow response will increase with time implies a hydrological response to a static change in basin state. However, the streamflow response to forest harvesting is time variant; the greatest impacts are immediate to short term, before forest regrowth returns vegetation to its previous state (Winkler *et al*., 2010). Thus, a different statistical approach is required to detect hydrological change from forest cover change, not a longer dataset as suggested by Swanson *et al*. (1986). The analysis conducted here compared the WY and monthly runoff ratios of disturbed to control sub-basin for the mean, peak, and low flows as well as differences in timing of peak, lower, median, and upper streamflow quartiles (Figure 11). This quantifies the runoff response from harvesting by normalizing it to the Middle Creek control sub-basin. This approach assumes that, apart from land cover changes, there is a fundamentally similar hydrological behaviour between adjacent sub-basins, an assumption supported by recent modelling studies of MCRB (Fang *et al*., 2013). Any

Table VI. Marmot Creek historical flood summary

Date	Peak streamflow (m^3/s)	Precipitation amount (mm)	Precipitation station	Distance from MCRB (km)
2013-06-20	2.8–3.3[a]	246	Upper Clearing, Fisera Ridge, and Hay Meadow	
1995-06-06	2.31	77.8	BGSI	14.49
1990-06-02	2.21	64.8	Mt Kidd	4.54
1971-06-06	2.18	79.5	Confluence 5	
1964-06-08	1.95	69.1	Boundary Ranger Station	5.87
1974-06-17	1.95	0	Confluence 5	
2005-06-18	1.92	149.2	BGSI	14.49

BGSI, Biogeosciences Institute; MCRB, Marmot Creek Research Basin.
[a] Estimated.

changes in the metrics between sub-basins immediately after the forestry operations are assumed to be due to forest harvesting. The only significant changes identified after disturbance were a trend for decreased peak flow at Cabin Creek (Figure 11c), more variable flow at Cabin Creek (Figure 11a), less variable flow at Twin Creek (Figure 11b), and a trend for earlier streamflow for Cabin Creek (Figure 11e). These differences make physical sense as large clear-cuts undergo rapid snow ablation, whereas small forest clearings undergo variable snow ablation rates that vary with slope and aspect (Musselman *et al*., 2008; Pomeroy *et al*., 2012; Ellis *et al*., 2013). The reduction in peak flow from the clear-cut sub-basin may be due to desynchronization of melt timing between clearing and forest (Pomeroy *et al*., 2012). The subtle differences noted in the Cabin Creek peak streamflow and timing of the peak flow did not translate into significant impacts for the greater Marmot Creek basin. Water yield and discharge timing at the basin scale were resilient to changes in forest cover at the basin scale.

Influence of teleconnections on hydrometeorology

A significant body of work has linked teleconnections, such as the PDO and ENSO, to variations in the hydrology and meteorology of Western Canadian river basins (Bonsal *et al*., 2001; Whitfield *et al*., 2010). To test for the impact of teleconnections on trends in Marmot Creek hydrology and driving meteorology, GLS regressions of trend, PDO, and ENSO terms were made with air temperature, precipitation, groundwater, snowpack, and streamflow observations at multiple elevations in MCRB and at BGSI. Table VII summarizes the results and show that the trend term describes the majority of the changes observed. Trend is a significant predictor for air temperature, low elevation precipitation, low elevation SWE, and groundwater away from streams. PDO is a significant predictor for peak SWE and mean water table at the well near the stream (Well 301), and mean air temperature and precipitation at BGSI. ENSO is only a significant predictor for low elevation precipitation in MCRB. The sign of these correlations agrees with

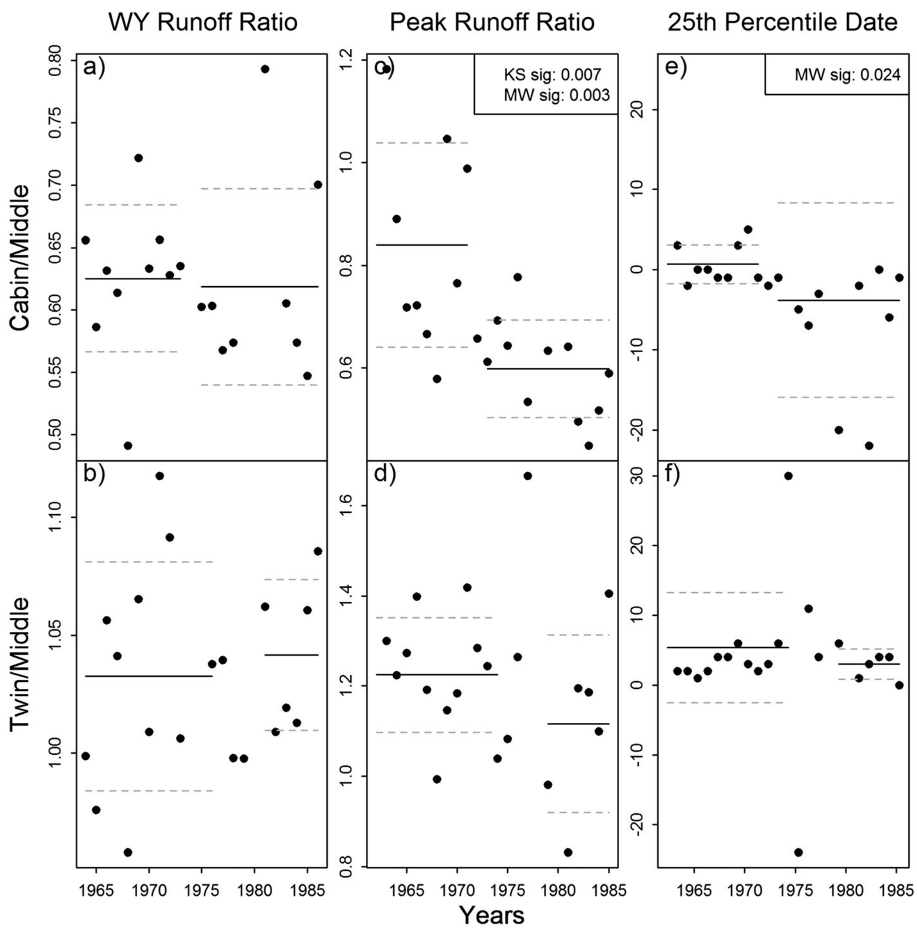

Figure 11. Pre-harvest and post-harvest Cabin Creek (a, c, and e) and Twin Creek (b, d, and f) water year (WY) runoff (a and b) and peak runoff (c and d) as ratio of Middle Creek runoff and difference in date (in days) of 25th percentile of flow (e and f). Mann–Whitney (MW) and Kolmogorov–Smirnov (KS) test significance values presented in plot legends. For the respective periods, the solid horizontal black lines are the mean values and the grey dashed lines are the standard deviations. Cabin Creek pre-harvest period was 1964–1973, and post-harvest period was 1975–1986. Twin Creek pre-harvest period was 1964–1976, and post-harvest period was 1980–1986. 1980 was removed from analysis because of Middle Creek gauging error

Table VII. Statistical association between MCRB hydrometeorological and groundwater variables and El Nino Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO)

	Variable	Sig. model terms	Data completeness[a] (%)	R^2
MCRB	May–Oct streamflow		98	0.08
TN1_FR	Annual Tmean	Trend	51	0.27
CN5_UC	Annual Tmean	Trend	51	0.46
TN1_FR	Annual_P		35	0.26
CN5_UC	Annual_P	ENSO,trend	31	0.34
Alpine	Peak SWE	PDO	53	0.20
Forest	Peak_SWE	PDO, trend	63	0.62
Well 305	Annual mean water table	Trend	37	0.43
Well 303	Annual mean water table	Trend	47	0.24
Well 301	Annual mean water table	PDO	51	0.37
BGSI	Annual Tmean	PDO, trend	89	0.34
BGSI	Annual P	PDO	91	0.15

BGSI, Biogeosciences Institute; MCRB, Marmot Creek Research Basin; CN5, Confluence 5; UC, Upper Clearing; TN1, Twin 1; FR, Fisera Ridge; SWE, snow water equivalent.
[a] Percentage of record that has observations: MCRB data completeness is with respect to 1963–2013 observation period, while BGSI data completeness is with respect to the 1939–2013 observation period.

previous studies (Gobena and Gan, 2006; Whitfield *et al.*, 2010), but the inconsistency of the results at MCRB is problematic as (1) air temperatures, lower and upper elevation groundwater (Wells 303 and 305), and lower elevation precipitation were not related to either PDO or ENSO; (2) coefficients of determination were very low for all PDO and ENSO models; and (3) the data completeness was low especially for SWE and groundwater data. The lack of a PDO–streamflow correlation, in spite of the SWE and groundwater PDO correlations, emphasizes resiliency in streamflow to teleconnections and the importance of long-term trends associated with changing climate in describing the variations observed at MCRB.

Resilience

The hydrology of MCRB is resilient to the changes in climate, forest disturbance, and extreme events experienced since 1962. Changes in climate have not had a great effect on upper elevation snow accumulation and melt that comprise the largest water input to the basin. The decrease in lower elevation snow accumulation has no measureable impact on basin-scale streamflow, as it may be compensated for by a subtle increase in precipitation at higher elevations. The lack of hydrological response to forest cover manipulations can be attributed to their limited area relative to the entire basin, the relatively low elevations of the disturbance, and the variations in slope and aspect enhancing desynchronizations in melt timing and so moderating the impact on peak streamflow generation (Ellis *et al.*, 2013). Subsurface processes that dominate streamflow generation may enhance resilience to extreme events such as floods by attenuating precipitation inputs. These subsurface processes also have an important role in providing resilience of MCRB to changes in climate and land cover owing to a large storage capacity. It is not clear whether the remarkable resilience of MCRB will continue as the climate changes. Climate change impacts are expected to increase at higher elevations with time; therefore, rainfall will have a greater role in streamflow generation processes as has already been observed at Reynolds Creek (Nayak *et al.*, 2010).

CONCLUSIONS

The meteorological, hydrological, and groundwater records at multiple elevations over 52 years in MCRB reveal remarkable hydrological resilience despite substantive changes in climate, extreme weather, and land cover. The hydrometeorology of the basin is changing, with air temperatures increasing at all elevations. The greatest air temperature increases are at low elevations and for winter daily minimums with smaller increases in summer and no change in spring. These results add an important elevational context to mountain trend analyses that have primarily used valley bottom meteorological stations (e.g. Zhang *et al.*, 2000). Although annual precipitation showed no trend, multiple day precipitation volume, interannual variability in precipitation, and spring precipitation volume have increased with the greatest increases at higher elevations. Low elevation snowpacks have declined dramatically, while higher elevation snowpacks show no change. Groundwater levels vary with proximity to the stream, and changes in snow accumulation and multiple day precipitation. A GLS regression analysis found that trend, rather than PDO or ENSO teleconnections, was the persistent term describing temporal changes in precipitation, temperature, groundwater level, and snow accumulation. No trends were identified in Marmot Creek streamflow volume, and magnitude or timing of peak flow from the establishment of the gauge in 1962 to its destruction in the flood of June 2013. If the streamflow gauge had been discontinued at the time of research funding cuts in 1987 or 2011, then apparent trends of −27% or −24%, respectively, would have been erroneously detected. This dataset shows the value of long-term monitoring and the importance of perseverance in hydrological observations.

Reanalysis of the Marmot Basin Project forest clearing experiment shows that there is no detectable change in water yield from sub-basins subjected to either large cut-block or small forest gap clearing. However, the cut-block treatment led to slightly earlier snowmelt timing and increased groundwater recharge. The differences associated with clear-cutting on the Cabin Creek sub-basin did not translate into measurable changes for the greater Marmot Creek basin, suggesting a larger scale hydrological resiliency to forest cover change. Observed hydrometeorological and groundwater changes in the MCRB are primarily dominated by an atmospheric forcing associated with gradual, but appreciable, climate change. In spite of climate change, the occurrence of an extreme precipitation event in 2013 and forest cover change MCRB shows muted hydrological responses at the basin scale. This suggests that storage and streamflow generation processes in this basin have attenuated the variability and trends observed in the meteorology and snow regime such that the basin-scale hydrology shows remarkable resilience to disturbance.

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REFERENCES

Adams PW, Flint AL, Fredriksen RL. 1991. Long-term patterns in soil moisture and revegetation after a clearcut of a Douglas-fir forest in Oregon. *Forest Ecology and Management* **41**: 249–263. DOI:10.1016/0378-1127(91)90107-7.

Alberta Environmental Protection. 1995. Oldman River Dam June 1995 Flood Summary Report. Water Resources Operations Division, Edmonton, Alberta.

Australian Bureau of Meteorology. 2015. S.O.I. (Southern Oscillation Index) archives – 1876 to present. http://www.bom.gov.au/climate/current/soihtm1.shtml (Accessed 9 July, 2015).

Beckstead G, Veldman W. 1985. *Marmot Creek experimental Watershed Study*. Water Survey of Canada: Calgary, Alberta.

Bonsal B, Shabbar A, Higuchi K. 2001. Impacts of low frequency variability modes on Canadian winter temperature. *International Journal of Climatology* **21**: 95–108.

Botter G, Basso S, Rodriguez-Iturbe I, Rinaldo A. 2013. Resilience of river flow regimes. Proceedings of the National Academy of Sciences of the United States of America **110**: 12925–12930. DOI: 10.1073/pnas.1311920110

Bronaugh D, Werner A. 2013. zyp package. R package version 2.2.

Creed IF, Gabor ZS, Buttle JM, Jones JA. 2011. Hydrological principles for sustainable management of forest ecosystems. *Hydrological Processes* . DOI:10.1002/hyp.8056.

Cunderlik JM, Ouarda T. 2009. Trends in the timing and magnitude of floods in Canada. *Journal of Hydrology* **375**: 471–480. DOI:10.1016/j.jhydrol.2009.06.050.

Davis RE, Lowit MB, Knappenberger PC, Legates DR. 1999. A climatology of snowfall–temperature relationships in Canada. *Journal of Geophysical Research* **104**: 11985–11994.

Dumanski S, Pomeroy JW, Westbrook CJ, 2015. Hydrological regime changes in a Canadian Prairie basin. *Hydrological Processes* DOI:10.1002/hyp.10567.

Ellis CR, Pomeroy JW, Brown T, MacDonald JP. 2010. Simulation of snow accumulation and melt in needleleaf forest environments. *Hydrology and Earth System Sciences* **14**: 925–940. DOI:10.5194/hess-14-925-2010.

Ellis CR, Pomeroy JW, Link TE. 2013. Modeling increases in snowmelt yield and desynchronization resulting from forest gap-thinning treatments in a northern mountain headwater basin. *Water Resources Research* **49**: . DOI:10.1002/wrcr.20089.

Fang X, Pomeroy JW, Ellis CR, MacDonald MK, DeBeer CM, Brown T. 2013. Multi-variable evaluation of hydrological model predictions for a headwater basin in the Canadian Rocky Mountains. *Hydrology and Earth System Sciences* **17**: 1635–1659. DOI:10.5194/hess-17-1635-2013.

Fisera Z. 1985. Snow accumulation in the Marmot Creek Research Basin – March 1985. Canadian Forestry Service Internal Report.

Fleming SW, Dahlke HE. 2014. Parabolic northern-hemisphere river flow teleconnections to El Nino-Southern Oscillation and the Arctic Oscillation. *Environmental Research Letters* **9**: DOI:10.1088/1748-9326/9/10/104007.

Fleming SW, Sauchyn DJ. 2013. Availability, volatility, stability, and teleconnectivity changes in prairie water supply from Canadian Rocky Mountain sources over the last millennium. *Water Resources Research* **49**: 64–74. DOI:10.1029/2012WR012831.

Gobena AK, Gan TY. 2006. Low frequency variability in southwestern Canadian stream flow: links with large-scale climate anomalies. *International Journal of Climatology* **26**: 1843–1869.

Golding DL, Swanson RH. 1978. Snow accumulation and melt in small forest openings in Alberta. *Canadian Journal of Forest Research* **8**: 380–388.

Granger RJ, Pomeroy JW. 1997. Sustainability of the western Canadian boreal forest under changing hydrological conditions – 2 – summer energy and water use. In *Sustainability of Water Resources under Increasing Uncertainty*, Rosjberg D, Boutayeb N, Gustard A, Kundzewicz Z. Rasmussen P (eds). IAHS Press: Institute of Hydrology, Wallingford, UK, No. **240**; 243–250.

Harder P, Pomeroy JW. 2014. Hydrological model uncertainty due to precipitation phase partitioning methods. *Hydrological Processes* . DOI:10.1002/hyp.10214.

Hayashi M, van der Kamp G. 2009. Progress in scientific studies of groundwater in the hydrologic cycle in Canada, 2003–2007. *Canadian Water Resources Journal* **34**: 177–186.

Hetherington ED, 1987. The importance of forests in the hydrological regime. In *Canadian Aquatics Resources Vol. 215*, Healey MC, Wallace RR (eds). Canadian Bulletin of Fisheries and Aquatic Sciences, Fisheries and Oceans Canada: Ottawa, Canada;179–211.

Hinkley DV. 1970. Inference about the change-point in a sequence of random variables. *Biometrika* **57**: 1–17.

Hood JL, Roy JW, Hayashi M. 2006. Importance of groundwater in the water balance of an alpine headwater lake. *Geophysical Research Letters* **33**: L13405. DOI: 10.1029/2006GL026611

Ihaka R, Gentleman R. 2011. R: a language for data analysis and graphics. *Journal of Computational and Graphical Statistics* **5**: 299–314.

Killick R, Eckley I. 2014. changepoint: an R package for changepoint analysis. *Journal of Statistical Software* **58**: 1–19.

Luckman BH. 1997. Developing a proxy climate record for the last 300 years in the Canadian Rockies – some problems and opportunities. *Climatic Change* **36**: 455–476.

Macdonald JS, Beaudry PG, Macisaac EA, Herunter HE. 2003. The effects of forest harvesting and best management practices on streamflow and suspended sediment concentrations during snowmelt in headwater streams in sub-boreal forests of British Columbia, Canada. *Canadian Journal of Forest Research* **33**: 1397–1407. DOI:10.1139/X03-110.

Mantua N. 2015. PDO Index. University of Washington Joint Institute for the Study of the Atmosphere and Ocean. http://research.jisao.washington.edu/pdo/PDO.latest (Accessed 9 July. 2015)

McLeod A. 2011. Kendall: Kendall rank correlation and Mann–Kendall trend test. R package version 2.2.

Mote PW, Hamlet AF, Clark MP, Lettenmaier DP. 2005. Declining mountain snowpack in western North America. *Bulletin of the American Meteorological Society* **86**: 39–49. DOI:10.1175/BAMS-86-1-39.

Musselman K, Molotch N, Brooks P. 2008. Effects of vegetation on snow accumulation and ablation in a mid-latitude sub-alpine forest. *Hydrological Processes* **22**: 2767–2776. DOI:10.1002/hyp.

Nayak A, Marks D, Chandler DG, Seyfried M. 2010. Long-term snow, climate, and streamflow trends at the Reynolds Creek Experimental Watershed, Owyhee Mountains, Idaho, United States. *Water Resources Research* **46**: 1–15. DOI:10.1029/2008WR007525.

Newton BW, Prowse TD, Bonsal BR. 2014. Evaluating the distribution of water resources in western Canada using synoptic climatology and selected teleconnections. Part 1: winter season. *Hydrological Processes* DOI:10.1002/hyp.10233.

Pennelly C, Reuter G, Flesch T. 2014. Verification of the WRF model for simulating heavy precipitation in Alberta. *Atmospheric Research* **136**: 172–192.

Peterson WT, Schwing FB. 2003. A new climate regime in northeast pacific ecosystems. *Geophysical Research Letters* **30**: 1896. DOI:10.1029/2003GL017528.

Peterson TJ, Western AW, Argent RM. 2012. Analytical methods for ecosystem resilience: a hydrological investigation. *Water Resources Research* **48**:W10531. DOI: 10.1029/2012WR012150.

Pinheiro J, Bates D, DebRoy S, Sarkar D, R, Development Core Team. 2012. nlme: linear and nonlinear mixed effects models. R package version 3. 1–105.

Pomeroy JW, Fang X, Ellis CR. 2012. Sensitivity of snowmelt hydrology in Marmot Creek, Alberta, to forest cover disturbance. *Hydrological Processes* **26**: 1891–1904. DOI:10.1002/hyp.9248.

Pomeroy JW, Fang X, Ellis CR, Guan M. 2011. Centre Report No. 10. Sensitivity of snowmelt hydrology on mountain slopes to forest cover disturbance. Centre for Hydrology, University of Saskatchewan; Saskatoon, SK.

Pomeroy JW, Granger RJ. 1997. Sustainability of the western Canadian boreal forest under changing hydrological conditions. I. Snow accumulation and ablation. In *Sustainability of Water Resources Under Increasing Uncertainty*, Rosjberg D, Boutayeb N, Gustard A, Kundzewicz Z. Rasmussen P (eds). IAHS Press: Institute of Hydrology, Wallingford, UK, No. **240**; 237–242.

Pomeroy JW, Gray DM. 1995. *Snow Accumulation, Relocation and Management*. Environment Canada: Saskatoon, SK.

Pomeroy JW, Marks D, Link T, Ellis C, Hardy J, Rowlands A, Granger RJ. 2009. The impact of coniferous forest temperature on incoming longwave radiation to melting snow. *Hydrological Processes* **23**: 2513–2525.

Pomeroy JW, Parviainen J, Hedstrom N, Gray DM. 1998. Coupled modelling of forest snow interception and sublimation. *Hydrological Processes* **12**: 2317–2337.

Pomeroy JW, Stewart R, Whitfield PH. in review. The 2013 flood event in the Bow and Oldman River basins: causes, assessment and damages. *Canadian Water Resources Journal* (in review).

Rood SB, Pan J, Gill K, Franks C, Samuelson GM, Shepherd A. 2008. Declining summer flows of Rocky Mountain rivers: changing seasonal hydrology and probable impacts on floodplain forests. *Journal of Hydrology* **349**: 397–410. DOI:10.1016/j.jhydrol.2007.11.012.

Rood SB, Samuelson GM, Weber JK, Wywrot KA. 2005. Twentieth-century decline in streamflows from the hydrographic apex of North America. *Journal of Hydrology* **306**: 215–233. DOI:10.1016/j.jhydrol.2004.09.010.

Sakamoto Y, Ishiguro M, Kitagawa G. 1986. *Akaike Information Criterion Statistics*. D. Reidel Publishing Company: Dordrecht/Tokyo.

Sen K. 1968. Estimates of the regression coefficient based on Kendall's Tau. *Journal of the American Statistical Association* **63**: 1379–1389.

Shook K, Pomeroy JW. 2012. Changes in the hydrological character of rainfall on the Canadian prairies. *Hydrological Processes* **26**: 1752–1766. DOI:10.1002/hyp.9383.

St. Jacques J, Sauchyn DJ, Zhao Y. 2010. Northern Rocky Mountain streamflow records: global warming trends, human impacts or natural variability? *Geophysical Research Letters* **37**: 1–5. DOI:10.1029/2009GL042045.

Stevenson DR. 1971. Collection and compilation of groundwater data in the Marmot Creek Experimental Basin. Groundwater Division, Research Council of Alberta.

Stewart IT, Cayan DR, Dettinger MD. 2004. Changes in snowmelt runoff timing in Western North America under a 'business as usual' climate change scenario. *Climatic Change* **62**: 217–232. DOI:10.1023/B:CLIM.0000013702.22656.e8.

Storr D. 1967. Precipitation variation in a small forested watershed, In *Western Snow Conference*. Boise, Idaho; 11–17.

Swanson RH, Golding DL. 1982. Snowpack management on Marmot watershed to increase late season streamflow, In *Western Snow Conference*: Reno, Nevada. 215–218.

Swanson RH, Golding DL, Rothwell RL, Bernier P. 1986. *Hydrologic Effects of Clear-Cutting at Marmot Creek and Streeter Watersheds*. Alberta, Forestry Service of Canada: Edmonton, Alberta.

Tennant CJ, Crosby BT, Godsey SE. 2015. Elevation-dependent responses of streamflow to climate warming. *Hydrological Processes* **29**: 991–1001. DOI:10.1002/hyp.10203.

Toop DC, de la Cruz NN. 2002. Hydrogeology of the Canmore Corridor and Northwestern Kananaskis Country, Alberta. Alberta Environment, Hydrogeology Section, Edmonton, Alberta; Report to Western Economic Partnership Agreement, Western Economic Diversification Canada.

Troendle CA, Kaufmann MR, Hamre RH, Winokur RP. 1987. Management of subalpine forests: building on 50 years of research, In *Proceedings of a Technical Conference*. Silver Creek, Colorado; 253.

Valeo C, Xiang Z, Bouchart FJ, Yeung P, Ryan MC. 2007. Climate change impacts in the Elbow River watershed. *Canadian Water Resources Journal* **32**: 285–302. DOI:10.4296/cwrj3204285.

Wheater HS, Gober P. 2013. Water security in the Canadian prairies: science and management challenges. *Philosophical Transactions of the Royal Society A* **371**: 20120409. DOI:10.1098/rsta.2012.0409.

Whitfield PH. 2012. Floods in future climates: a review. *Journal of Flood Risk Management* **5**: 336–365. DOI:10.1111/j.1753-318X.2012.01150.x.

Whitfield PH. 2014. Climate station analysis and fitness for purpose assessment of 3053600 Kananaskis, Alberta. *Atmosphere-Ocean* 1–21. DOI:10.1080/07055900.2014.946388.

Whitfield PH, Moore RD, Fleming SW, Zawadzki A. 2010. Pacific decadal oscillation and the hydroclimatology of Western Canada – review and prospects. *Canadian Water Resources Journal* **35**: 1–28. DOI:10.4296/cwrj3501001.

Wilcox R, Schönbrodt F. 2013. The WRS package for robust statistics in R. R package version 0.20.

Winkler RD, Moore RD, Redding TE, Spittlehouse DL, Smerdon BD, Carlyle-Moses DE. 2010. The effects of forest disturbance on hydrologic processes and watershed response. In *Compendium of Forest Hydrology and Geomorphology in British Columbia*, Pike RG, Redding TE, Moore RD, Winker RD, Bladon KD (eds). B.C. Min. For. Range, For. Sci. Prog., Victoria, B.C. and FORREX Forum for Research and Extension in Natural Resources: Kamloops, B.C.; 179–212.

Woo, M-K, Pomeroy, JW. 2011. Snow and runoff: processes, sensitivity and vulnerability. In *Changing Cold Environments: A Canadian Perspective*, French H, Slaymaker O (eds). John Wiley & Sons, Ltd: Chichester, UK. DOI: 10.1002/9781119950172.ch6.

Yue S, Pilon P. 2004. A comparison of the power of the t test, Mann–Kendall and bootstrap tests for trend detection. *Hydrological Sciences Journal* **49**: 21–37.

Yue S, Pilon P, Cavadias G. 2002a. Power of the Mann–Kendall and Spearman's Rho tests for detecting monotonic trends in hydrological series. *Journal of Hydrology* **259**: 254–271. DOI:10.1016/S0022-1694(01)00594-7.

Yue S, Pilon P, Phinney B, Cavadias G. 2002b. The influence of autocorrelation on the ability to detect trend in hydrological series. *Hydrological Processes* **16**: 1807–1829. DOI:10.1002/hyp.1095.

Zhang X, Vincent LA, Hogg WD, Niitsoo A. 2000. Temperature and precipitation trends in Canada during the 20th century. *Climate Research* **38**: 395–429.

Zheng A, Basher R, Thompson C. 1997. Trend detection in regional-mean temperature series: maximum, minimum, mean, diurnal range, and SST. *Journal of Climate* **10**: 317–326.

