Future streamflow simulation in a snow-dominated Rocky Mountain headwater catchment

Ram P. Neupane, Jan F. Adamowski, Joseph D. White and Sandeep Kumar

ABSTRACT

The Rocky Mountains in North America are comprised of headwater snow catchments that provide sustained seasonal flow downstream. Changes in streamflow over the last half century in these basins may be associated with changing climate with increased temperature and variable precipitation, shifting seasonal hydrology. We investigated potential changes in future hydrology in a Rocky Mountain headwater catchment by simulating water budgets of the Athabasca River located in Jasper National Park, Canada. Potential hydrologic changes were predicted using a calibrated version of the Soil and Water Assessment Tool (SWAT). Future discharge and other parts of the catchment water budget were projected based on the global circulation model (GCM) derived from the Special Report on Emission Scenarios (SRES) for the latter part of the century (2081–2099). A projected decrease in future precipitation resulted in reduced mean annual streamflow, by up to 86%, compared to the baseline period for the catchment. Projected summer streamflow decreased from 58 to 39%. Streamflow increased from 13 to 26% during the spring, dampening the dominance of summer peak-flow hydrology. Colder winters for the future scenarios increase the overall proportion of precipitation as winter snowfall. However, dramatically lower precipitation estimated for this basin will drive water limits for the future.

Key words | climate change, discharge, headwater, lower precipitation, scenario, SWAT

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INTRODUCTION

Mountain systems are sensitive to climate change (Kohler & Maselli 2009) due to temperature and precipitation effects on snowpack development and, subsequently, snowmelt derived water supply (Barnett *et al.* 2005). The snow hydrology of mountains, primarily in the form of snowpack and glacier formation and loss, regulate freshwater that supplies the world's major river systems including the Indus, Ganges-Brahmaputra, and Colorado (Beniston *et al.* 1997; Water 2009; Ficklin *et al.* 2013). The impacts of climate variability on water resources can affect areas ranging from several hundred square kilometers (e.g. Milly *et al.* 2005; Lubini & Adamowski 2013) to as little as 8 km^2 (e.g. Wang *et al.* 2008). Climate change impacts on hydrologic processes in mountains has resulted in significant interest from scientists, and resource managers, and decision-making bodies due to concerns regarding changes in the reliability of water supplies (Li *et al.* 2013).

Projected changes in temperature and precipitation may influence the mean hydrologic processes of river basins that can affect the frequency and magnitude of extreme hydrologic events (Praskievicz & Chang 2009). Sustainable future water management of mountainous basins relies on accurate representation of climate variables where 50–70% of the total precipitation may fall in the form of snow (Serreze *et al.* 1999), and the seasonal snowmelt of the spring and early summer may account for 50–80% of the total annual runoff (Stewart *et al.* 2004).

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The headwater basins in the Rocky Mountains of Canada are critical for freshwater resources. However, for the past century, river flows for these basins have declined by an average of 0.22%/yr (Rood *et al.* 2005). The reason for these declines is unclear, however snow and ice accumulation and melt timing are suspected as mechanisms affected by increased temperature and changes in precipitation (Hamlet & Lettenmaier 1999). Within the Canadian Rockies, glacier area has already been reduced by 25%, attributed to climate change beginning in the late 1800s (Luckman & Kavanagh 2000). The mean annual temperature has been increasing in the upper elevation of these mountain systems since the 1950s (Luckman 1990), with reduced snowpacks and shifts in seasonal release of meltwater from these basins (Barnett *et al.* 2005; Lapp *et al.* 2005). Changes in the low-order headwater basins of the Canadian Rocky Mountains may not be reflective of overall basin water budgets (Peters *et al.* 2013). However, precipitation changes in the region coupled with the ecologically sensitive nature of headwater catchments warrants investigation into possible future shifts in hydrology.

Simulated climate change effects on geographically low-elevation agriculture-dominated watersheds have shown changes in overall water supply rates, potentially affecting the reliability of flows during times of high human usage (e.g. Chien *et al.* 2013; Novotna *et al.* 2014; Neupane & Kumar 2015). In headwater basins of the North American Rocky Mountains, climate change impacts on hydrological processes have been generally shown to be associated with an earlier onset of melting (White *et al.* 1998; Cayan *et al.* 2001; Mote *et al.* 2005; Stewart 2009; Tinkham *et al.* 2015), and decreased mean annual streamflow (Zhang *et al.* 2001; Rood *et al.* 2005). In this study, we simulated the streamflow of the Rocky Mountain watershed with detailed incorporation of snow/glacier data using a process-based hydrologic model (i.e. Soil and Water Assessment Tool, SWAT). We then estimated the effects of potential climate variability on key hydrological processes including precipitation and snowmelt in the study watershed. This study explores the regional hydrologic response to climate change, in view of the impacts on ecosystem-services and the oil sands industry under a range of climate projections. These estimations may be important to assess the timing and source of future water availability, with the emphasis on expected changes related to sustained streamflow and potential ecosystem functioning in the watershed.

MATERIALS AND METHODS

Study watershed

For this study, we focused on the headwater watershed of the Athabasca River, referred to as the Upper Athabasca watershed, located in south-western Canada. This watershed covers an area of about 3,500 km^2, with an elevation range from 922 to 3,736 m above mean sea level (Figure 1). The Athabasca River is the second largest river in Alberta, originating from the Athabasca glacier of the Rocky Mountains located in Jasper National Park that flows north-east emptying into Lake Athabasca downstream. The watershed is part of the Mackenzie River that eventually discharges into the Arctic Ocean. The watershed downstream is critical for diverse ecosystems including a staging area for a large number of waterfowl, primarily during spring and autumn seasons, and is recognized internationally as a RAMSAR wetland and UNESCO World Heritage site (Schindler *et al.* 2007; Pavelsky & Smith 2009).

Geologically, the watershed area is underlain almost entirely by sedimentary bedrock units ranging in age from Proterozoic to early Tertiary, with large variations in snow/glacier processes that cause landslides in the region (Jackson 2002; Selkowitz *et al.* 2002a). The climate of the western part of this region is mainly controlled by a stronger maritime influence originated from the North Pacific Ocean. The eastern part has a distinctly more continental climate (Selkowitz *et al.* 2002b). The annual temperature in the Lake Athabasca region for the period of 1971–2000 ranged between −3.5 and 7.6 °C with a mean value of 2.1 °C (http://climate.weather.gc.ca/climate_normals/index_e.html). Precipitation varies dramatically within the span of low and high elevation ranges. Mean annual precipitation is ~504 mm, primarily in the form of snowfall, with higher precipitation in the north-eastern part of the region. These variations in climatic factors over relatively small distances affect microclimate, vegetation distribution, and ecosystem services of the region (Peterson *et al.* 1997).

Precipitation and stream discharge of the Athabasca River basin are shown in Figure 2. Analysis of 23 years

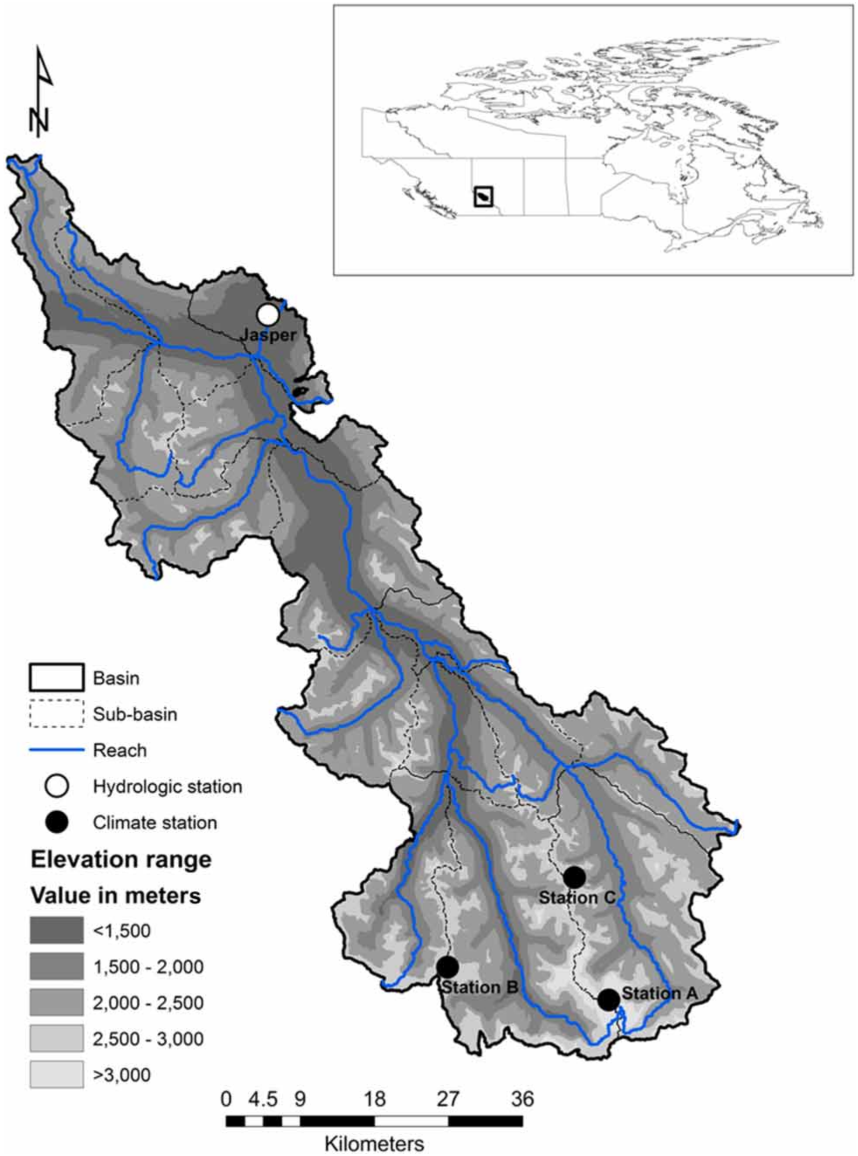

Figure 1 | The Athabasca River watershed located in south-western Canada, including stream network and spatial hydro-meteorological locations used for model simulations in this study.

(1979–2001) of precipitation data for eight different spatial locations of the basin (http://rda.ucar.edu/pub/cfsr.html) showed a higher amount of precipitation from May to August (Figure 2(a)). The minimum precipitation amount of 76 mm occurred in February and the maximum precipitation of 153 mm was observed in June. Stream discharge measured from May to September at Jasper in the Athabasca River basin (common outlet shown in Figure 1) showed a maximum value of 253 m^3/s in July (http://wateroffice.ec.gc.ca/search/searchResult_e.html) potentially due to higher summer precipitation that increased surface runoff in these months. Seasonal analysis of precipitation and stream discharge data indicated that about 25% of total annual precipitation in the basin occurred during the spring season which corresponded to 11% of the mean annual stream discharge in the same season (Figure 2(b)). The minimum precipitation contribution of 20% occurred during the winter season, corresponding to a minimum discharge contribution of 4% during the same season. Water resources were most abundant during summer months

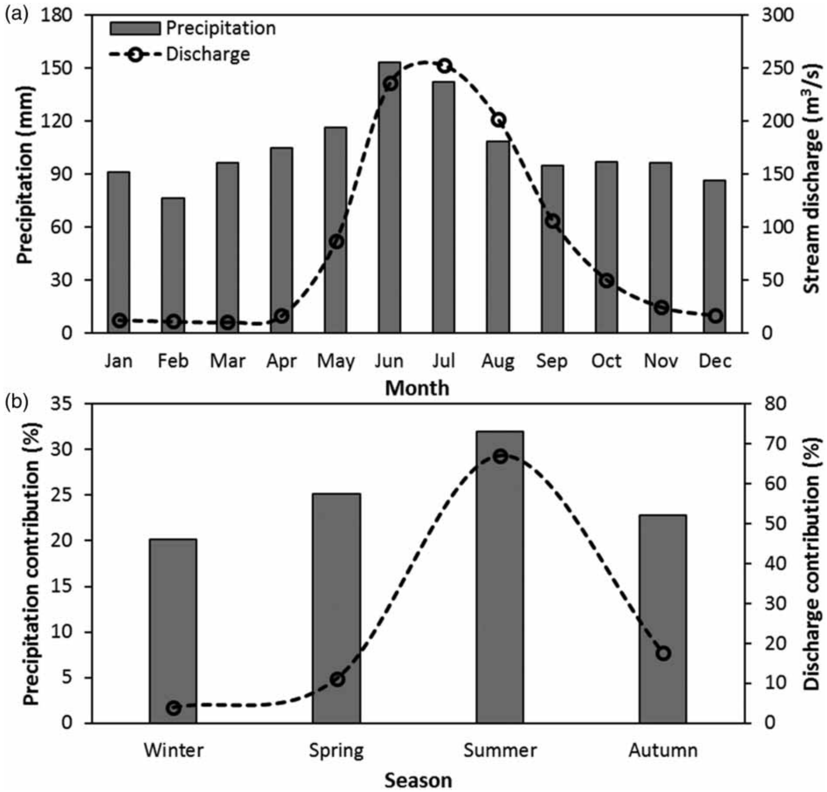

Figure 2 | Hydrographs for the Athabasca River watershed: (a) mean monthly precipitation and stream discharge and (b) seasonal precipitation and stream discharge contribution. The 23 years (1979–2001) of precipitation data obtained from the National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR) and 31 years (1971–2001) of stream discharge data obtained from Canadian Hydrometric station number 07AA002 were used to derive these hydrographs.

with precipitation and snow/glacier-melt generally exceeding potential evapotranspiration and, therefore, the maximum discharge contribution of 67% occurred during this season.

Hydrologic model

Assessing climate change impacts on potential hydrologic processes of mountain basins is complicated due to climatic heterogeneity, lack of hydro-meteorological data, and uncertainty in snow/ice characteristics (Beniston 2003). There are various models developed for modeling watershed hydrologic processes subject to solid and liquid precipitation, including commercial software such as MIKE-SHE (http://mikebydhi.com) and public domain models such as HBV (Bergström 1992), Xinanjiang Model (Zhao *et al.* 1995), HEC-HMS (Azmat *et al.* 2016), SRM (Vafakhah *et al.* 2015), and SWAT (Neitsch *et al.* 2009). SWAT, a process-based hydrologic and water quality model developed for the USDA Agricultural Research Service (ARS), has gained international acceptance as a robust interdisciplinary model with a user-friendly interface set-up in a GIS framework that can integrate multiple environmental processes for development of better-informed policy decisions (Gassman *et al.* 2005). The model was originally developed to predict the impact of agricultural land management practices on water, sediment, and agricultural chemical yields in large-sized ungauged basins (Arnold *et al.* 1995; Neitsch *et al.* 2009). It has also been successfully calibrated and used to estimate the effects of potential climate variability on hydrologic processes for a number of diverse global basins (Stonefelt *et al.* 2000; Eckhardt & Ulbrich 2003; Song & Zhang 2012; Neupane *et al.* 2014, 2015; Awan *et al.* 2016).

Briefly, SWAT simulations are based on first dividing the entire watershed into sub-basins. Next, hydrologic response units (HRUs) are defined based on unique and user-defined combinations of land use, soil type, and topographic slope.

Within each HRU, a water budget is calculated based on the storage within snowpack, soil, shallow aquifer, and a deep aquifer. Streamflow, or the total amount of water leaving the HRU and entering into the main channel (*WYLD* in mm) each day, is calculated by:

$$WYLD = SURQ + LATQ + GWQ - TLOSS \quad (1)$$

where *SURQ* is the surface runoff (mm), *LATQ* is the lateral flow contribution to stream discharge (mm), *GWQ* is the groundwater contribution to stream discharge (mm), and *TLOSS* is the transmission losses from the watershed system (mm) (Arnold *et al.* 2011).

Snowmelt is modeled using a temperature index with air temperature, snowpack temperature, melt rate, and a measure of areal coverage of snow as inputs (Fontaine *et al.* 2002; Neitsch *et al.* 2009). Each day, the amount of snow melted is computed using (all presented in mm of H_2O):

$$SNO = SNO + R_{day} - E_{sub} - SNO_{mlt} \quad (2)$$

where *SNO* is the total amount of water in the snowpack on a given day, R_{day} is the amount of precipitation, E_{sub} is the amount of sublimation, and SNO_{mlt} is the amount of snowmelt. Orographic effects on temperature and snow accumulations are accounted in SWAT by defining elevation bands as simulation units.

Glacier contribution to meltwater is not explicitly modeled by the current SWAT version. However, because of its importance to stream discharge in our study watershed, we modeled glacier meltwater and added it to post-simulation stream discharge. Glacier melt was simulated by a simple degree-day model (Hock *et al.* 2005; Singh *et al.* 2006) in which melting ice was related to air temperature:

$$M = \begin{cases} D.(T_{av} - T_{gmlt}), & \text{when } T_{av} > T_{gmlt} \\ 0, & \text{otherwise} \end{cases} \quad (3)$$

where M is the depth of melt water (mm/day), D is the degree-day factor for ice melt (mm/day°C), T_{av} is the mean daily air temperature (°C), and T_{gmlt} is the threshold value for ice melt (°C). Seasonal changes of glacier melt were defined by a sinusoidal function (Neitsch *et al.* 2009) and calculated by:

$$D = \frac{b_{gmlt,6} + b_{gmlt,12}}{2} + \frac{b_{gmlt,6} - b_{gmlt,12}}{2}.sin\left[\frac{2\pi}{365}(t - 81)\right] \quad (4)$$

where $b_{gmlt,6}$ is the melt factor for June 21st (mm H_2O day^{-1} °C), $b_{gmlt,12}$ is the melt factor for December 21st (mm H_2O day^{-1} °C), and t is the day of the year. The ice volume accumulated in glaciated regions of the basin was assessed based on an empirical relation between surface area and mean thickness of the ice as mentioned in Liu *et al.* (2003).

A simple model was used to estimate stream water temperature from the SWAT simulations using weighted values derived from the contribution of different water sources to the stream water yield (Equation (1)). For each component, temperatures were assigned based on water sources. Rainfall was assumed to be the daily average temperature. Groundwater was assumed to be the annual mean temperature (≈4 °C) and snow/glacial meltwater was assumed to be 0.1 °C.

Input spatial data

Spatial data required for SWAT simulations included topography, land use, and soil properties. Topographic data were acquired from the global digital elevation model (GDEM) sourced from the Advanced Spaceborne Thermal Emission and Reflection Radiometer (http://gdem.ersdac.jspacesystems.or.jp/search.jsp) with a 30 × 30 m resolution. The GDEM data were then used to delineate the watershed boundary, stream network, and topographic characteristics such as terrain length and slope of the stream channels. We used classified multispectral data derived from the Environmental Satellite (ENVISAT) Medium Resolution Imaging Spectrometer (MERIS) (GlobCover, http://due.esrin.esa.int/page_globcover.php) with a 300 × 300 m spatial resolution as land use input into the model (Arino *et al.* 2009). Based on the inclusion of this land use data, about 7% of the total basin area is covered by permanent snow/ice (Figure 3(a)).

The only soil data available for this basin were from the Food and Agriculture Organization of the United Nations (FAO 1995; Reynolds *et al.* 1999) with a broad spatial resolution of 10 km. To refine soil characteristics similar to the

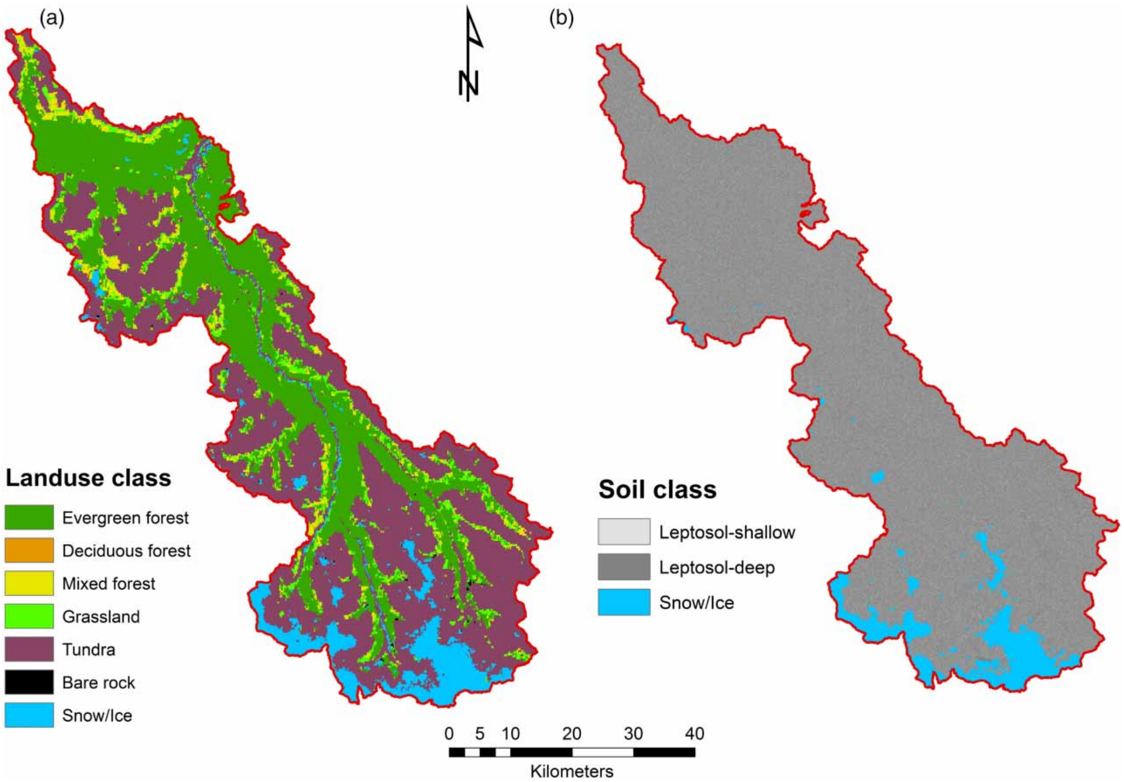

Figure 3 | Different land use class and soil distribution of the Athabasca River watershed. The soil is shown with two different depth classes (shallow and deep layers).

grain size of the topographic and landcover data, we first calculated the topographic saturation index (TSI) (Neupane *et al.* 2015) from the GDEM data using the flow accumulation and slope functions in ArcGIS. We then defined two soil layers (shallow soil with <10 cm from surface and deep soil >10 cm from surface; Figure 3(b)) from the TSI data using the maximum soil depth found from the original FAO soil data to help scale our TSI values (Neupane *et al.* 2014). Other important soil characteristics such as bulk density and nutrient characteristics were used 'as is' with values input from a user soil database based on the FAO/UNESCO Soil Map of the World (FAO/UNESCO 2003).

When soil data were combined with land use and slope values, we derived a total of 291 HRUs based on minimum area threshold values of 5, 10, and 10% for each land use, soil, and slope categories, respectively. From the land cover data, permanent glaciers were defined in nine HRUs from three different sub-basins with a total area of 236 km^2 and a total volume of 31 km^3. To account for topographic gradient, we defined elevation bands in every 500 m elevation within each sub-basin (Fontaine *et al.* 2002). For this, the mean elevation of each elevation band and percentage of the sub-basin area within that band were entered as the SWAT model has the ability to include up to 10 elevation bands within each sub-basin. Details of the elevation gradient in each sub-basin with the number of elevation bands are shown in Figure 4.

Hydro-meteorological data

For model calibration and confirmation, we used 23 years (1979–2001) of daily precipitation and daily maximum and minimum temperature data derived from grid-based observed meteorological data. The data were obtained from the National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR) (http://rda.ucar.edu/pub/cfsr.html) for eight different spatial locations and the details are given in Table 1. These point data represent an area of ~38 km^2 and have been successfully used for hydrologic simulations in other watershed systems (Fuka *et al.* 2013;

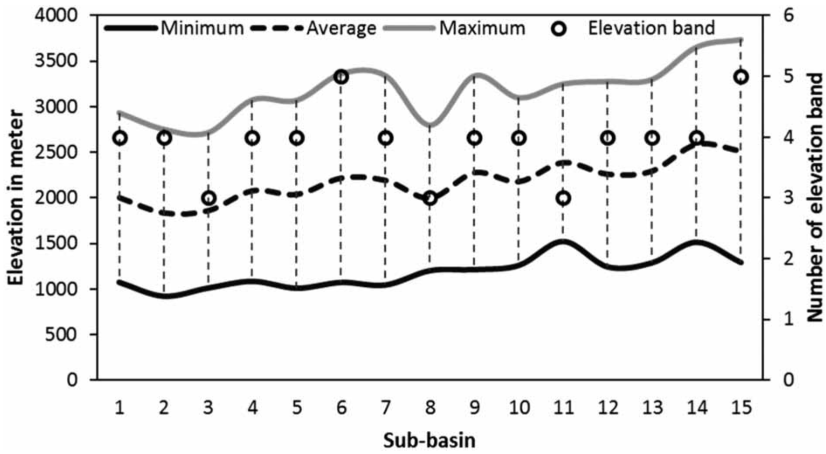

Figure 4 | Total number of sub-basins with their elevation range and number of elevation bands derived for SWAT simulation of precipitation and temperature in the Athabasca River watershed.

Table 1 | The hydro-meteorological stations with their spatial locations used in this study

Station	Type	Latitude (°)	Longitude (°)	Elevation (a.m.s.l.)
Jasper	Hydrology	52.91	−118.06	1,021
526-1175	Meteorology	52.61	−117.50	2,757
526-1178	Meteorology	52.61	−117.81	1,422
526-1181	Meteorology	52.61	−118.13	2,198
526-1184	Meteorology	52.61	−118.44	1,734
529-1175	Meteorology	52.92	−117.50	2,017
529-1178	Meteorology	52.92	−117.81	2,251
529-1181	Meteorology	52.92	−118.13	1,205
529-1184	Meteorology	52.92	−118.44	1,764
Station A	Meteorology	52.20	−117.34	3,312
Station B	Meteorology	52.22	−117.63	2,706
Station C	Meteorology	52.33	−117.42	2,686

Note: The meteorological stations presented in numbers are grid-based numbers for each location obtained from National Centers for Environmental Prediction; a.m.s.l. is above mean sea level.

Dile & Srinivasan 2014). Relative humidity, solar radiation, and wind speed data were calculated using a weather generator input (*.wgn*) file incorporated within SWAT using daily measured temperature and precipitation data. To account for glacier melt data, three spatial locations (Stations A, B, and C, as shown in Figure 1) were established at high elevations of glaciated regions to simulate mean daily air temperature applying the mountain climate simulator (MT-CLIM) (www.ntsg.umt.edu/project/mtclim) for the period 1982–2001 using daily meteorological inputs derived from the CFSR station grid-number 526-1178 and incorporated with the degree-day model. The derived glacier melt data were then incorporated with the SWAT model simulation outputs of stream discharge to compare with further simulations. Temperature and precipitation lapse rates required by SWAT for the distribution of meteorological variables based on elevation bands within each sub-basin were derived from previous studies conducted in US mountain systems (Fontaine *et al.* 2002; Pradhanang *et al.* 2011). Daily and monthly measured stream discharge data taken from the Canadian Hydrometric station number 07AA002 (http://wateroffice.ec.gc.ca/search/searchResult_e.html) located at Jasper (Figure 1) were compared with model simulation outputs.

Calibration, confirmation, and sensitivity analyses

Stream discharge simulated for the period 1979–1991 using SWAT default parameter values was first compared with measured stream discharge data taken from the gauging station at Jasper to analyze the un-calibrated model performance, referred to as the pre-calibration simulation. Subsequent simulations were organized as calibration (1979–1991) and confirmation (1992–2001) periods. The initial three years of calibrated simulation outputs were disregarded as a model spin-up period that allows the model to cycle multiple times to minimize the effect of user estimated parameter values (Zhang *et al.* 2007). For calibration and confirmation analysis of the model, we applied the SWAT-CUP (Abbaspour et al. 2007; Abbaspour 2012). The 95% prediction uncertainty (95PPU) (*P* factor) and the thickness of 95PPU (*R*-factor) were used to evaluate the accuracy of calibration and uncertainty analysis (Rostamian et al. 2008; Arnold et al. 2012). The Sequential Uncertainty Fitting (SUFI-2) algorithm, a semi-automatic inverse modeling approach, was used for this study due to its better capability of handling a large number of parameters with a lower number of model runs (Yang et al. 2008).

The global sensitivity analysis integrated within SUFI-2 was used to test 21 SWAT hydrologic parameters for stream discharge simulation in parallel with the calibration procedure. The derived new parameter values from calibration and confirmation analyses were incorporated with the SWAT database for further simulations. To study the influence of elevation bands as controls of orographic precipitation and temperature induced-melt, two calibration and confirmation simulation scenarios were run: (1) elevation bands were included and glacier melt data were not included (band-no glac) and (2) both elevation band and glacier melt data were included (band-glac). Calibrated model outputs were considered as the baseline data, representing current hydrologic conditions of the watershed that were compared with further simulation outputs. The Nash–Sutcliffe efficiency (NSE), Percent Bias (PBIAS), root mean square error (RMSE), and coefficient of determination (r^2) were used to assess the goodness of fit of the model simulation outputs (Moriasi et al. 2007; Golmohammadi et al. 2014):

$$\mathrm{NSE} = 1 - \frac{\sum (Y^{obs} - Y^{sim})^2}{\sum (Y^{obs} - Y^{mean})^2} \tag{5}$$

$$\mathrm{PBIAS} = \frac{\sum (Y^{obs} - Y^{sim})}{\sum Y^{obs}} \times 100 \tag{6}$$

$$\mathrm{RMSE} = \sqrt{\frac{1}{N}\sum_{i=1}^{N} (Y^{sim} - Y^{obs})^2} \tag{7}$$

$$\mathrm{r}^2 = \frac{(\mathrm{n}\sum Y^{obs}Y^{sim} - \sum Y^{obs}\sum Y^{sim})^2}{\left[n\sum (Y^{obs})^2 - (\sum Y^{obs})^2\right]\left[n\sum (Y^{sim})^2 - (\sum Y^{sim})^2\right]} \tag{8}$$

where Y^{obs} is the measured data, Y^{sim} is the model simulation output, Y^{mean} is the mean of measured data, and n is the number of observations in the period under consideration.

Potential climate change scenarios

For this study, we chose the B1 (low), A1B (medium), and A2 (high) emission scenarios as the representative of all extreme conditions expected for the 21st century that are known as the Special Report on Emission Scenarios (SRES) developed by the Intergovernmental Panel on Climate Change (IPCC) (Maurer et al. 2010). These scenarios include both natural and anthropogenic drivers of climate change. We used daily Bias-Correction Constructed Analogue (BCCA) average temperature and precipitation data estimated for the SRES (http://gdo-dcp.ucllnl.org/downscaled_cmip_projections/) (Maurer et al. 2010; Brekke et al. 2013) for the end of the 21st century. These data are downscaled at 1/8 degree (~12 × 12 km) spatial resolution that are suitable for hydrologic assessment studies. The bias-correction follows a basic approach of smoothening monthly mean values to avoid abrupt discontinuity to compensate for dry months. This generally helps to narrow the differences to obtain the best fit model. The general circulation model (GCM) structure is a major source of uncertainty for estimating the hydrologic impacts (Kay et al. 2009; Bennett et al. 2012); however, the Coupled Model Intercomparison Project Phase 3 (CMIP3) multimodel dataset (Meehl et al. 2007) were used in our study, based on their wider applicability with better performance, and more specifically over the western part of North America (Werner 2011). Also, these datasets were selected based on the hydrological impact studies recently conducted at the Pacific Climate Impacts Consortium (PCIC) for multiple watersheds of western Canada (Shrestha et al. 2012; Schnorbus et al. 2014). The average temperature and precipitation data derived from eight GCMs were incorporated into the SWAT for estimating their effects on annual and seasonal hydrologic processes including precipitation, surface runoff, stream discharge, water yield, evapotranspiration, soil water content, snowfall, and snowmelt of the Athabasca River basin.

To study the potential hydrologic changes, specifically for the end of the 21st century, the model simulations were run for the period 2081–2099 on a monthly basis. However, we ignored the initial three years (2081–2083) of simulation outputs for our analysis as a model spin-up period to compare with the baseline simulation results.

RESULTS AND DISCUSSION

Model parameterization, calibration, and confirmation

For this study, the key hydrologic parameters used for the Athabasca River basin SWAT simulations were selected based on the information derived from previous studies conducted in complex snow-dominated mountain basins (e.g. Pradhanang *et al.* 2011; Neupane *et al.* 2014, 2015) (Table 2). Model calibration and confirmation analyses are presented in Table 3. The correlation between measured and simulated discharge values during calibration and confirmation were improved when snow and glacier melt were specifically incorporated. The model showed the highest correlation with observed data for monthly simulations for the calibration of the band-glac scenario with NSE = 0.92, PBIAS = −14, RMSE = 27, and $r^2 = 0.94$. The statistical values were 0.85, −9, 34, and 0.87 for these indices, respectively for the band-glac scenario for the simulation confirmation period. These higher correlation indices indicate the importance of including glacier processes for predicting stream discharge simulation in the study watershed, similar to the findings of Abbaspour *et al.* (2010). Those researchers found that the highest correlation was obtained with the addition of glacier melt water, though their simulated area was larger than that used in our study. Rahman *et al.* (2013) also showed a similar high correlation between measured and model simulated stream discharge in the complex glacier-dominated upper Rhone River watershed specifically by the addition of elevation

Table 2 | Sensitivity results with the ranking of key SWAT parameters for stream discharge in the Athabasca River watershed including the range of parameter values adopted from Muleta & Nicklow (2005), Rostamian *et al.* (2008), Pradhanang *et al.* (2011), and Neupane *et al.* (2014, 2015)

Parameter	Description	Range	Optimal value	t-stat	Rank
v_SMFMX	Maximum melt rate for snow during the year (mm/°C-day)	0–10	0.54	18.017	1
v_SMTMP	Snow melt base temperature (°C)	(-5)–5	-4.26	18.009	2
r_CN2	Surface runoff curve number for moisture condition II	(-0.40)–0.40	0.21	8.760	3
v_TLAPS	Temperature lapse rate (°C/Km)	(-7)–(-5)	-7.00	7.193	4
v_SFTMP	Snowfall temperature (°C)	(-5)–5	3.16	6.821	5
v_TIMP	Snow pack temperature lag factor	0.01–1	0.09	5.387	6
v_RCHRG_DP	Deep aquifer percolation fraction	0–1	0.60	3.041	7
v_OV_N	Manning's n value for overland flow	0–0.80	0.43	2.807	8
v_ALPHA_BNK	Baseflow alpha factor for bank storage (days)	0.05–1	0.96	2.605	9
v_GW_REVAP	Groundwater 'revap' coefficient	0.02–0.20	0.15	2.080	10
v_CH_K2	Effective hydaulic conductivity in main channel alluvium (mm/hr)	0–150	83.02	1.803	11
v_SMFMN	Minimum melt rate for snow during the year (mm/°C-day)	0–10	0.08	1.291	12
v_GW_DELAY	Groundwater delay time (days)	0–400	120.00	1.144	13
v_ALPHA_BF	Baseflow alpha factor (days)	0–1	0.75	0.912	14
v_PLAPS	Precipitation lapse rate (mm H_2O/km)	12.11–37.77	24.18	0.874	15
v_CH_N2	Manning's n value for the main channel	0–0.30	0.08	0.820	16
v_REVAPMN	Threshold depth of water in the shallow aquifer for revap (mm H_2O)	0–100	22.65	0.653	17
v_EPCO	Plant uptake compensation factor	0.001–1	0.39	0.136	18
v_GWQMN	Threshold depth of water in the shallow aquifer required for return flow to occur (mm H_2O)	0–100	80.55	0.026	19
v_ESCO	Soil evaporation compensation factor	0.001–1	0.90	0.025	20
v_SURLAG	Surface runoff lag time (days)	1–24	8.83	0.023	21

Table 3 | Model performance statistics for pre-calibration, calibration, and confirmation simulations at both daily and monthly time periods (no band-no glac: elevation bands and glacier melt data were not used; band-no glac: elevation bands were used and glacier melt data were not used; band-glac: both elevation band and glacier melt data were used)

	Pre-calibration		Calibration				Confirmation			
	no band-no glac		band-no glac		band-glac		band-no glac		band-glac	
Statistics	**Daily**	**Monthly**	**Daily**	**Monthly**	**Daily**	**Monthly**	**Daily**	**Monthly**	**Daily**	**Monthly**
NSE	−1.78	−0.34	0.65	0.90	0.68	0.92	0.53	0.80	0.56	0.85
PBIAS	−46	−46	−12	−12	−13	−14	−11	−11	−12	−9
RMSE	176	111	62	30	60	27	66	40	63	34
r^2	0.28	0.49	0.68	0.93	0.70	0.94	0.60	0.81	0.62	0.87

bands and glacier melt data in the basin; however, they were focused on a relatively larger glacier basin considering only nine hydrologic parameters for calibration and confirmation of the model.

Improved statistical values between measured and model simulated stream discharge data were also clearly represented by the hydrographs obtained after model calibration (Figure 5). However, the model showed overestimation during peak summer flows for daily simulations, potentially due to higher glacier melt estimation combined with measurement errors occurring during the same season (Rossi *et al.* 2009). The model also overestimated the winter baseflow components for both daily and monthly simulations that may be attributed to the difficulty in accounting sub-surface flow contribution to stream discharge, including a lack of proper representation for near stream saturation associated with excess runoff (Larose

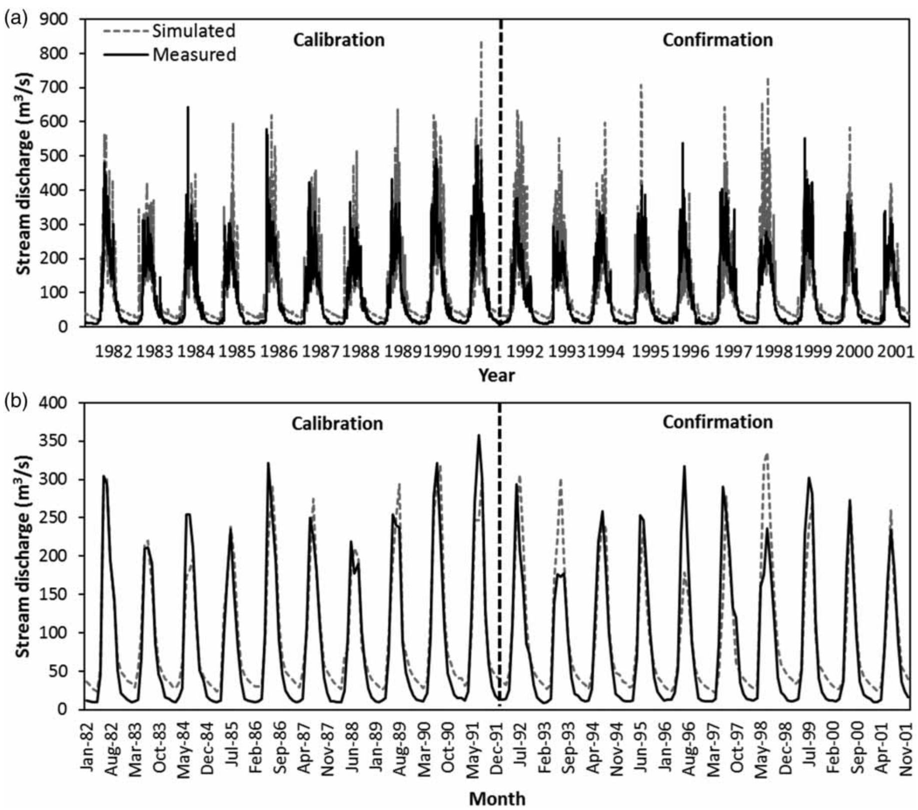

Figure 5 | Hydrographs obtained during model calibration (1982–1991) and confirmation (1992–2001) periods: (a) daily and (b) monthly time step.

et al. 2007). Additional research is needed to address these discrepancies between measured and model simulated outputs, especially important for assessing seasonal changes in discharge.

In Rocky Mountain basins, the ablation of glaciers intensifies in the summer and this together with higher snowmelt at high elevations prolongs the higher flows during this season (Woo & Thorne 2003). There are only a limited number of studies conducted in parts of the North American Rocky Mountain system regarding changes in snowmelt hydrology (Ahl *et al.* 2008; Abbaspour *et al.* 2010; Watson & Putz 2013). However, ours is one of the first to identify the impacts of climate change at a smaller spatial scale.

Model sensitivity and uncertainty analyses

The most sensitive input parameters regarding stream discharge simulation were identified on the basis of global sensitivity analysis, and are presented in Table 2. This analysis identified that the five most sensitive parameters for this study were SMFMX (maximum snowmelt rate), SMTMP (snowmelt base temperature), CN2 (surface runoff curve number), TLAPS (temperature lapse rate), and SFTMP (snowfall temperature), out of which four were observed as snow-related parameters. Betrie *et al.* (2015), in a study conducted for the entire Athabasca basin, also found the same parameters as the most important for SWAT simulations, however with different calibrated values reflective of the scale differences. The calibrated lower SMFMX value of 0.54 likely indicates minimal melting of snow at the beginning of the summer season in the basin. The SMTMP variable is the threshold temperature above which snowmelt occurs and therefore influences the simulated hydrographs' shape and peak flows. The negative calibrated SMTMP value indicates an early start of the melting process. We found the TLAPS as another influencing variable to influence the accuracy of simulated stream discharge. This is likely due to the extreme topography of the basin, similar to other Rocky Mountain watersheds affecting adiabatic lapse rates over short geographic distances influencing the amount and type of precipitation (Minder *et al.* 2010). The higher negative TLAPS value of $-7.0\,^{\circ}$C/km (Table 2) may indicate the influence of local climatic factors such as the presence of higher moisture-bearing winds in the basin coming from the Pacific Ocean. The CN2 variable is a runoff coefficient obtained by calculating the amount of surface runoff following a precipitation event and assigned to each HRU based on land use, soil type, and moisture content. As surface runoff is extremely sensitive to CN2, higher values increase surface runoff, reduce the infiltration rate, and decrease the groundwater recharge (Singh *et al.* 2005). The EPCO (plant uptake compensation factor), GWQMN (threshold depth of water in shallow aquifer required for return flow), ESCO (soil evaporation compensation factor), and SURLAG (surface runoff lag time) variables were found to be the least sensitive for simulating stream discharge. Uncertainty analysis of simulated stream

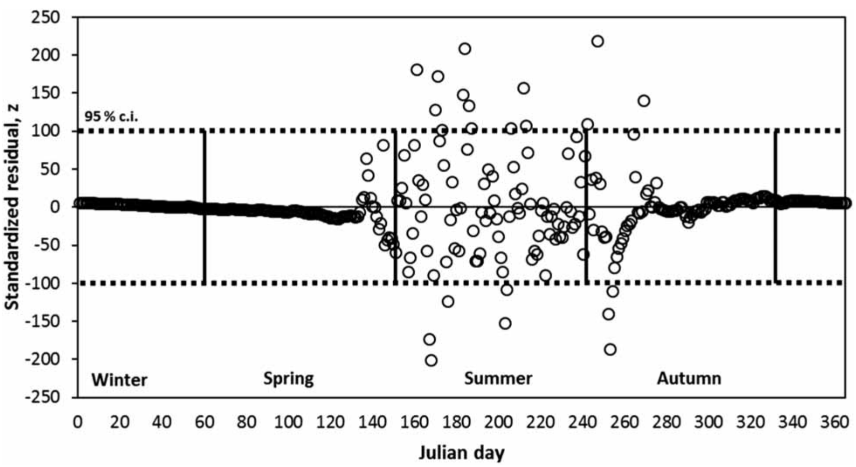

Figure 6 | Standardized residual error (z-scores) for SWAT model simulated stream discharge (m^3/s) *vs* measured data for the Athabasca River watershed (*Note:* the data were taken from the calibration period of 1982).

discharge data also showed significant over- and underestimation values, mainly during the summer season (Figure 6), that may be attributed to measurement errors occurring during high flow seasons (Rossi *et al.* 2009).

Projected annual changes in hydrologic processes of the basin

Based on our simulation results, the effects of potential changes in climatic variables, specifically temperature and precipitation, are likely to bring substantial changes to the discharge regime of the Athabasca River. The most important results of this study indicate that streamflow for the Athabasca River in Jasper National Park may be 86% less (Figure 7(c)) than current flows by the end of this century, caused by climate change. The reason for this decline is attributed to decreased precipitation with minor effects associated with changes in snowmelt. The projected changes from the GCMs used for this study are not extremes, rather averages of 16 different models. These results support a continuing trend of lower water budgets for this catchment that appears to have begun in the 1950s at an approximate rate of −0.22%

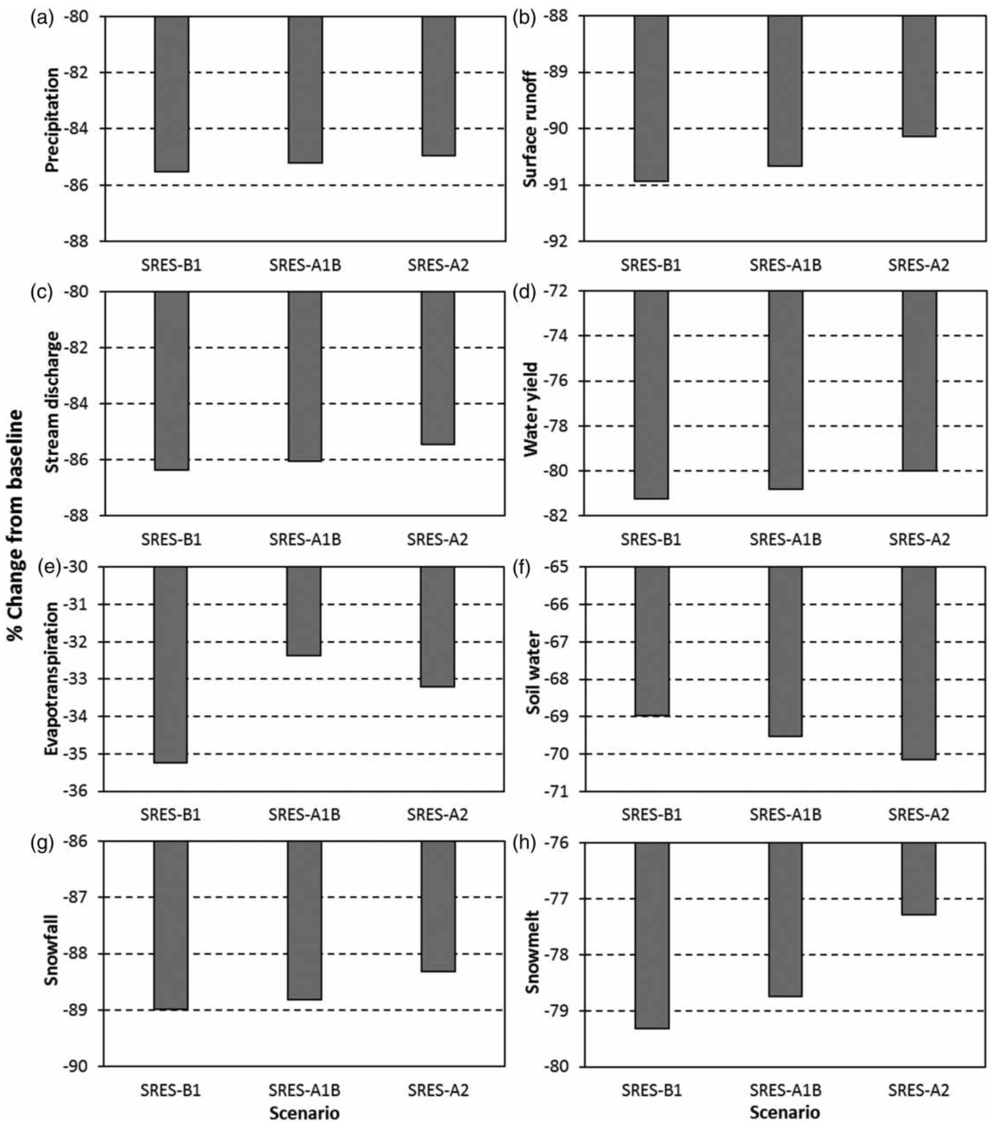

Figure 7 | Mean annual changes of (a) precipitation, (b) surface runoff, (c) stream discharge, (d) total water yield, (e) evapotranspiration, (f) soil water content, (g) snowfall, and (h) snowmelt from the baseline simulation for potential climate change scenarios.

per year (Rood *et al.* 2005; Peters *et al.* 2013). Our results, based on the predicted 86% less discharge for the 2080s, indicated a decline of nearly 1.09% per year.

Monthly and annual changes in temperature and precipitation derived for B1, A1B, and A2 scenarios for the Athabasca River basin are presented in Table 4. Mean annual temperature was estimated to be higher than the current value with values ranging from 0.9 to 2.5 °C. The mean monthly temperature varied substantially, ranging from −2.7 to +5.2 °C. We found future precipitation decreased in all scenarios with a maximum change of approximately −86% in SRES-B1 (Figure 7(a)). In addition, maximum precipitation change was projected during the summer months and minimal change during the winter months. Lower precipitation corresponded with reduced surface runoff by 91% in the same scenario (Figure 7(b)). The SRES-B1 scenario had 35% lower ET than the baseline (Figure 7(e)) associated with a coincident reduction in the soil water (Figure 7(f)). The SRES-A2 scenario had the highest change in soil water of −70% compared to the baseline, likely as a result of increased temperature with resultant effects on latent heat exchange.

Higher temperatures combined with lower precipitation in the basin resulted in less snowfall and snowmelt for all scenarios with maximum changes of −89% (Figure 7(g)) and −79% (Figure 7(h)), respectively, for the SRES-B1 scenario. Future lower precipitation combined with reduced meltwater contribution will affect future water availability (Kundzewicz *et al.* 2007), exasperating current conflicts in human water use and ecosystem services in this drainage basin (Schindler & Donahue 2006; Squires *et al.* 2009).

Projected seasonal changes in hydrologic processes of the basin

Future seasonal changes in hydrologic processes of the Athabasca River catchment are shown in Figure 8. At present, the seasonal partitioning of contributed precipitation in the basin were estimated to be 19, 25, 33, and 23% for the winter, spring, summer, and autumn seasons, respectively (Figure 8(a)). Shifts in precipitation inputs based on the GCM data show, for example, that for the SRES-A2 scenario, more precipitation is expected in winter and autumn seasons of 29 and 27%, respectively, and less summer precipitation, 24%, as compared with the current climate. For the SRES-B1scenario, spring precipitation decreased from 25 to 19% for the basin.

Despite a lower precipitation contribution during the spring season, we estimated substantially higher surface runoff during this season in all scenarios, with a maximum contribution up to 47% in the SRES-A2 scenario (Figure 8(b)). This corresponded to a higher spring snowmelt contribution of 60% in the same scenario (Figure 8(h)). Lower summer surface runoff of 48% in SRES-A2 corresponded to minimal precipitation, leading to a reduced stream discharge contribution of 39% in the same scenario (Figure 8(c)). Due to potential higher spring surface runoff, total water yield of the basin also increased in all the scenarios up to 28% in SRES-A2 (Figure 8(d)). Similarly, reduced precipitation coupled with earlier spring snowmelt also reduced the water yield substantially during the summer season in all scenarios

Table 4 | Projected scenarios for the Athabasca River basin for future hydrologic assessments: (a) mean annual and monthly temperature change (°C) and (b) mean annual and monthly cumulative precipitation change (%) (*Note:* the average GCM data were derived from the period between 2081 and 2099 to use as climate forcing data for simulation modeling)

Scenario	Month												
	1	2	3	4	5	6	7	8	9	10	11	12	Annual
(a) Temperature													
SRES-B1	−2.7	1.0	0.5	1.7	1.8	2.4	2.6	2.4	1.5	1.5	−0.3	−1.5	0.9
SRES-A1B	−1.6	1.3	1.2	2.3	2.6	3.7	4.1	3.8	2.9	2.5	0.2	−0.6	1.9
SRES-A2	−1.4	1.8	1.4	2.8	2.9	4.2	5.2	5.0	4.1	3.3	0.9	−0.1	2.5
(b) Precipitation													
SRES-B1	−30	−38	−60	−65	−56	−59	−53	−49	−43	−43	−38	−26	−47
SRES-A1B	−27	−35	−58	−62	−56	−59	−60	−53	−43	−37	−34	−17	−45
SRES-A2	−22	−34	−55	−61	−54	−61	−65	−55	−41	−35	−29	−10	−44

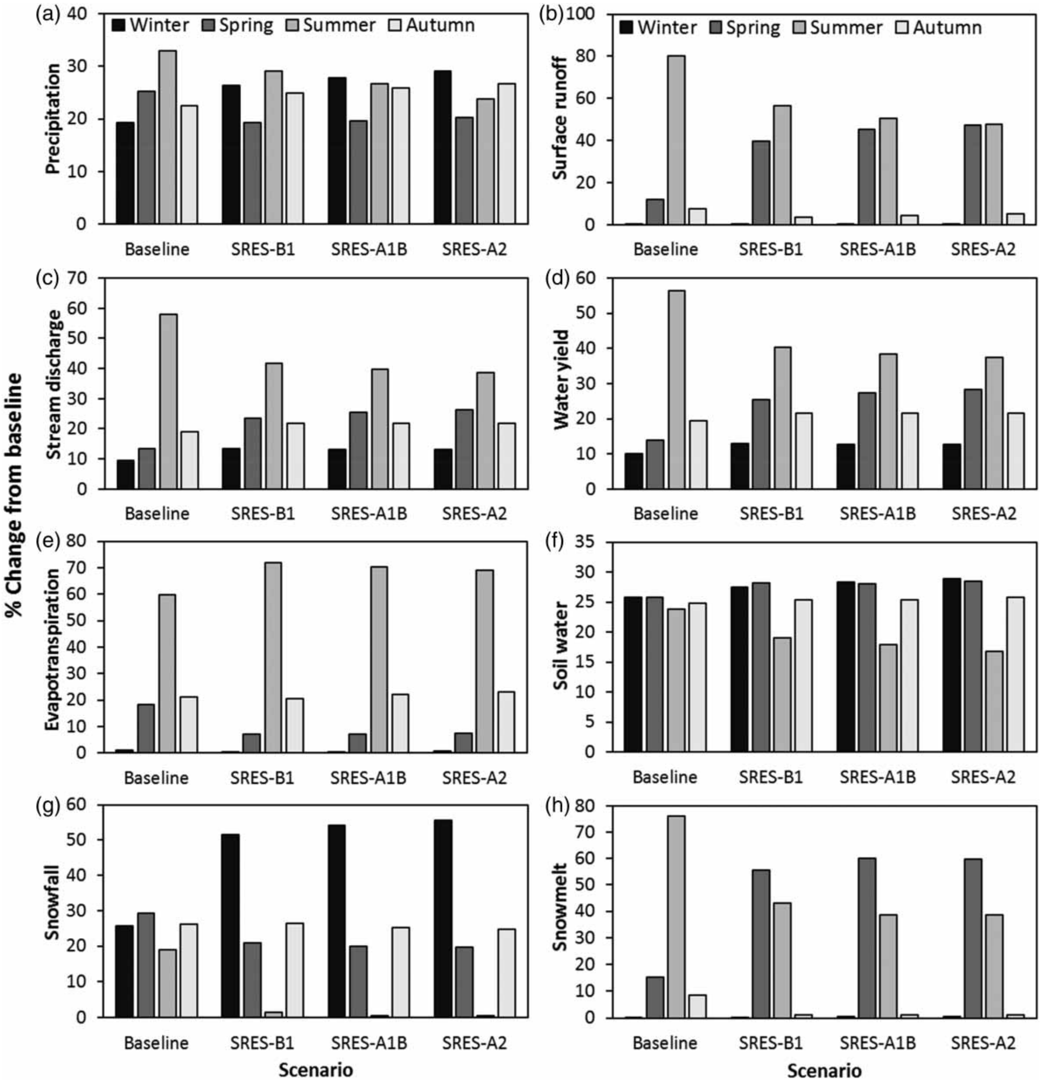

Figure 8 | Potential seasonal changes in (a) precipitation, (b) surface runoff, (c) stream discharge, (d) total water yield, (e) evapotranspiration, (f) soil water content, (g) snowfall, and (h) snowmelt simulated for different climate change scenarios of the Athabasca River watershed (Winter: December–February, Spring: March–May, Summer: June–August, and Autumn: September–November).

with a minimum contribution of 37% in SRES-A2. Due to changes in temperature and precipitation, projected snowfall in the basin also changed with a substantially lower contribution from melting during the summer season (Figure 8(g)). Therefore, the summer season was projected to be more dry with higher evapotranspiration (Figure 8(e)) and lower soil water content (Figure 8(f)). Climate change impact studies in recent years have begun for regional scale estimations. For example, Barnett *et al.* (2008) evaluated climate change impacts on water supply and regional hydrology in the western part of the United States using a high-resolution model and, similar to our estimations, they assessed increased runoff during spring season and a significant reduction during summer months at the peak of water demands.

For the baseline simulation, we estimated a maximum summer contribution of 58% for total mean annual stream discharge (Figure 8(c)). This flow includes the current higher summer precipitation combined with a higher glacier contribution during summer months in the basin. The large meltwater proportion is also due to abundant current snowpack levels in addition to seasonal glacier ablation from the high elevation of the Rocky Mountains, primarily during the summer season (Woo & Thorne 2003). However, this hydrologic peak during the summer is likely to diminish as glaciers in these basins continue to rapidly retreat in response to ongoing climate warming (Marshall & White 2010; Grover *et al.* 2014). Our simulations show decreased summer discharge with increased flows during winter and spring seasons in the future. This corresponded to the findings of Shepherd *et al.* (2010) who investigated similar seasonal changes in future water availability of Oldman River Basin in the North American Rocky Mountains.

Stream water temperature estimates for this study indicated that changes in water sources and potential increase in air temperature affected summer values most, with the largest difference between baseline estimated at 5.0 °C and the SRES-B1 scenario at 7.3 °C (Figure 9). For the SRES-A2 scenario, autumn stream water temperatures were elevated compared with all other simulations.

Climate change impacts on future water availability

Lower streamflow and projected changes in seasonal flow caused by lower precipitation and a potential increase in temperature affects the snowpack of this Rocky Mountain basin and this will directly limit future water availability for any anticipated commercial appropriation, such as water for oil sands development or municipal water in this region (Pavelsky & Smith 2008; Squires *et al.* 2009). The shifts in quantity and timing of hydrologic discharge were estimated for this snow-dominated region over the coming

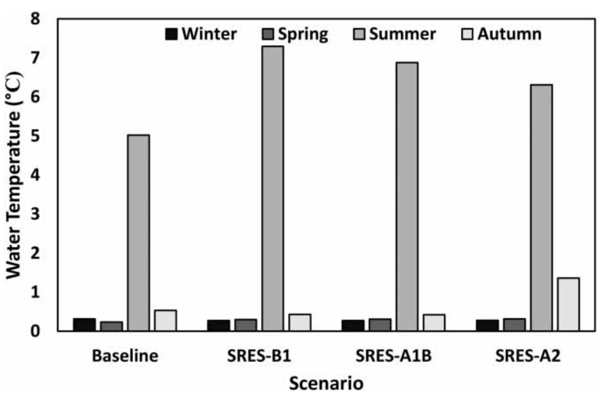

Figure 9 | Estimated seasonal stream temperature (°C) of the Athabasca River based on seasonal contributing water sources from the catchment for the baseline and 2081–2099 climate change scenarios.

century and which may influence freshwater ecosystem-services. Stream water quantity and flow timing influence critical water characteristics such as flow rate, dissolved chemical constituents, and water temperature affecting aquatic invertebrate and vertebrate life cycles (Fagre *et al.* 1997; Hauer *et al.* 2007). As water temperatures have been increasing at higher elevations in similar Rocky Mountain catchment reaches, sensitive aquatic invertebrate taxa may be running out of space with local extirpations, resulting in trophic shifts in mountain streams (Giersch *et al.* 2015).

Overall, warming is expected for the basin with higher temperature values significantly affecting both the water quality and quantity of the region (Chmura 2005). Water temperatures predicted here are crude estimates and do not include important factors such as energy balances associated with temperature effects on river ice or river hydrodynamics affecting ice formation and breakup. However, because changes in source and temperature increases during spring seasons are likely to occur, ice breakup in the river and subsequent flooding are likely to occur earlier in the season. Changes in seasonal hydrology can affect riparian ecosystems by changing the timing and frequency of flooding events (de Rham *et al.* 2008). These potential changes in temperature and precipitation combined with anthropogenic disturbance may increase the impacts on hydrologic components and surrounding ecosystem-services of the basin (Schindler & Donahue 2006).

Future hydrologic estimation in complex mountainous basins is complicated due to large climatic variations, and it is further elevated by the uncertainties of the General Circulation Models (GCMs) that are too coarse to adequately represent the orographic details of the mountains (Beniston 2003). Also, our research is based on a 'static' impact approach that does not include mesoclimatic interactions such as those from Pacific Decadal Oscillation, Arctic Oscillation, and the North Atlantic Oscillation which have been shown, in this basin, to greatly modify the timing and amount of streamflow (Burn 2008). While our simulations can be improved by using Representative Concentration Pathways (RCPs) laid under IPCC AR5 (IPCC 2013), these results should facilitate the consideration of implications for the possibility of future water limitations in this catchment and deliberation of climate change adaptation strategies.

CONCLUSIONS

The SWAT model was calibrated and tested for the snow-dominated upper Athabasca River watershed and the simulation results indicate that the model is a useful tool for assessing the effects of potential changes in temperature and precipitation on hydrological processes in the basin. Potential discharge of the Athabasca River was found to be dramatically lower for the 2080–2099 period, corresponding with lower precipitation, consequently leading to a reduced total annual water availability in the basin. Seasonally, winter discharge of the river was estimated to be higher, corresponding to similar changes in precipitation. Despite reduced future spring precipitation, discharge of the river was estimated to be higher, potentially indicating the importance of higher spring snowmelt in the basin. However, we estimated a substantial reduction in water availability in the basin during the summer season when there is a higher demand for water resources for agricultural, industrial, and domestic sectors. Finally, these assessments may improve our understanding of hydrological consequences of potential changes in temperature and precipitation of complex snow-dominated mountain basins, and provide better knowledge for future water resources management. Better representation of snowmelt processes through field-scale experiments should be investigated to refine model performance, primarily for simulating winter season base flow.

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