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The effect of climate change on the simulated streamflow of six Canadian rivers based on the CanRCM4 regional climate model

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Abstract

The effect of climate change is investigated on the hydro-climatology of six major Canadian rivers (Mackenzie, Yukon, Columbia, Fraser, Nelson, and St. Lawrence), in particular streamflow, by analyzing results from the historical and future simulations (RCP 4.5 and 8.5 scenarios) performed with the Canadian regional climate model (CanRCM4). Streamflow is obtained by routing runoff using river networks at 0.5° resolution. Of these six rivers, Nelson and St. Lawrence are the most regulated. As a result, the streamflow at the mouth of these rivers shows very little seasonality. Additionally, the Great Lakes significantly dampen the seasonality of streamflow for the St. Lawrence River. Mean annual precipitation (P), evaporation (E), runoff (R), and temperature increase for all six river basins considered and the increases are higher for the more fossil fuel-intensive RCP 8.5 scenario. The only exception is the Nelson River basin for which the simulated runoff increases are extremely small. The hydrological response of these rivers to climate warming is characterized by their existing climate states. The northerly Mackenzie and Yukon River basins show a decrease in evaporation ratio (E/P) and an increase in runoff ratio (R/P) since the increase in precipitation is more than enough to offset the increase in evaporation associated with increasing temperature. For the southerly Fraser and Columbia River basins, the E/P ratio increases, and the R/P ratio decreases due to an already milder climate in the Pacific north-western region. The seasonality of simulated monthly streamflow is also more affected for the southerly Fraser and Columbia Rivers than for the northerly Mackenzie and Yukon Rivers as snow amounts decrease and snowmelt occurs earlier. The streamflow seasonality for the Mackenzie and Yukon rivers is still dominated by snowmelt at the end of the century even in the RCP 8.5 scenario. The simulated streamflow regime for the Fraser and Columbia Rivers shifts from a snow-dominated to a hybrid/rainfall-dominated regime towards the end of this century in the RCP 8.5 scenario. While we expect the climate change signal from CanRCM4 to be higher than other climate models, owing to the higher-than-average climate sensitivity of its parent global climate model, the results presented here provide a consistent overview of hydrological changes across six major Canadian river basins in response to a warmer climate.

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1. Introduction

As the global population and the standard of living increases so does the strain on freshwater resources. The natural availability of water is determined by the balance between precipitation (P) and evaporation (E) (although the term evapotranspiration is more correct). When precipitation exceeds evaporation, which is determined primarily by available energy, the water that doesn't evaporate or transpire (either at the surface or after infiltration into the soil) termed runoff (R) is carried by the rivers to the oceans. The seasonality of precipitation, its partitioning into snow and rainfall, and the seasonality of snowmelt and evaporation, all of which are determined by the climate in a given catchment or a river basin eventually determine the seasonality of runoff. As anthropogenic climate change progresses, changes in the mean annual amounts and the seasonality of these different water budget components will lead to corresponding changes in runoff (Trenberth et al., 2007). Changes in precipitation extremes are also expected to lead to corresponding changes in the extremes of streamflow. The changes in streamflow have implications for floods and power generation. While runoff is expressed in similar units to precipitation and evaporation (depth of water per unit time, e.g. mm/s or m/year), streamflow is the volume of water generated per unit time and requires multiplication with the area over which runoff is generated. Streamflow is also routed down the river network which introduces a time lag and attenuation of the peak runoff. As a result, the streamflow is expressed in units of volume per unit time (e.g. m^3/s or $km^3/year$).

Output from climate and Earth system models (ESMs) remains the primary source of information for evaluating climate change impacts. Current approaches that rely on information generated by ESMs, to obtain an estimate of how future streamflow may potentially change, may

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be classified into two broad categories. The first approach uses simulated runoff directly from the land surface component of single or multiple climate models which may be routed downstream to obtain streamflow at the mouths of river basins and at different points along a given river network (Miller and Russell, 1992; Arora and Boer, 2001; Zhang et al., 2014). Using direct runoff output from climate models has the benefit that the calculated changes in runoff are physically consistent with the altered radiative balance of the Earth in response to increases in the concentrations of greenhouse gases (GHGs). The corresponding changes in the general circulation of the atmosphere result in the associated changes in near-surface temperature, precipitation, and the hydrological cycle. This approach suffers from three limitations – 1) the biases in the climate simulated by the climate model, 2) the fact that the land surface components of climate models are not calibrated for a given river basin but rather designed to operate in a reasonably realistic way over the whole globe, and 3) the coarse resolution of global climate models (GCMs). The last limitation is partially addressed when data from finer-resolution regional climate models is used. The biases in the simulated climate do affect the simulated runoff for the current climate. Despite this, however, the response to climate change is relatively robust and there is useful information in the simulated change that can be used to inform adaptation measures. The second approach attempts to overcome these limitations by downscaling and/or bias-correcting climate from climate models for future scenarios and uses that to drive a well-calibrated hydrological model for given catchments or river basins (Ismail et al., 2020; Yoosefdoost et al., 2022; Gosling et al., 2011; Miller et al., 2021). The second approach is more prevalent for watershed to regional scale impacts and adaptation studies. Given the large effort involved in downscaling and bias-correcting raw climate data from climate models, most

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current impact studies use downscaled and bias-corrected data put together by other groups rather than specifically doing this for their project. Recent examples include the downscaled and bias-corrected climate data for the conterminous United States (Thrasher et al., 2013) based on climate model output from the fifth phase of the Coupled Model Intercomparison Project (CMIP5), and statistically downscaled and bias-corrected data based from five CMIP5 models, available at the global scale, tailored to the requirements of the Inter-Sectoral Impact Model Intercomparison Project (ISIMIP) (Lange, 2019). Both these datasets have found large applications in the impacts and adaptation community. The processes of downscaling and bias correction are distinct, and they both have their inherent limitations. There are several examples of the limited ability of bias-correction to correct and to downscale variability, that bias-correction can potentially cause implausible climate change signals, and there remain uncertainties, substantial contradictions, and sensitivity to assumptions between the different downscaling methods (Maraun et al., 2017; Maraun, 2016; Hewitson et al., 2014). Well-calibrated hydrological models are generally suitable for a given catchment or a river basin but their application has not been extended to large-scale global or regional hydrologic modelling studies since it is typically not straightforward to tune model parameters for all river basins. In the end, both approaches are complementary to each other.

Future hydrologic projections using the second approach (hydrological modes driven by statistically downscaled and bias-adjusted climate models) are available for selected river basins in Canada. The results over the Prairies and British Columbia (Shrestha et al., 2021b; Sobie and Murdock, 2022) generally indicate shorter snow cover duration, earlier snowmelt, and reduced annual maximum snow water equivalent as the climate warms. Streamflow projections across

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Canada generally indicate earlier occurring snowmelt-driven peak flow, increased winter flow, and decreased summer flow (MacDonald et al., 2018; Shrestha et al., 2019; Islam et al., 2019; Dibike et al., 2021; Budhathoki et al., 2022). Annual streamflow is generally projected to increase, with higher increases in the northern basins (Bonsal et al., 2020; Stadnyk et al., 2021). However, these projections are based on different climate and hydrological models, downscaling methods, emissions scenarios, and future periods, and no consistent set of projections is available across all major river basins of Canada.

In this study, we have used the first approach to provide a consistent set of projections across all major river basins of Canada. We investigate the effect of climate change on the annual, monthly, and daily streamflow characteristics of six major Canadian rivers (Mackenzie, Yukon, Columbia, Fraser, Nelson, and St. Lawrence) using runoff output from simulations performed with version 4 of the Canadian Regional Climate Model (CanRCM4). The river basins of the Yukon and Columbia Rivers cover part of the United States of America as well. We used daily runoff generated from CanRCM4 for the historical period and for the two future scenarios (representative concentration pathways (RCP) 4.5 and 8.5). The spatial resolution of runoff data from CanRCM4 is 0.22° which is equivalent to about 12 km at 60° N (Canada lies between approximately 42°N and 83°N). Additionally, we utilized a large ensemble (50 realizations) of the CanRCM4 (CanRCM4-LE) at 0.44° resolution to quantify uncertainties associated with internal variability. We then routed this runoff through river networks at 0.5° resolution to evaluate streamflow at the mouths of major Canadian rivers. The Mackenzie, Yukon, and Fraser Rivers are somewhat less regulated than the heavily regulated Nelson, Columbia, and St. Lawrence Rivers. The routing scheme used here does not take into account dams and reservoirs and therefore the

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modelled streamflow represents natural streamflow. This aspect is discussed in more detail in Section 2.0

2. Models and data

Equation (1) summarizes the water balance over a given grid cell or a river basin for a given timescale.

$$P = E + R + \Delta S \qquad (1)$$

where ΔS is the change in water storage including that in soil moisture, snow, and the canopy water storage combined. When a system is in equilibrium, at annual or longer timescales $\Delta S = 0$ and $P = E + R$. ΔS, however, may not be zero even over long timescales when a system is not in equilibrium e.g., when snow is accumulating or is melting consistently. We evaluated P, E, and R components of equation (1) simulated by CanRCM4 for each of the six river basins, considered in this analysis, and routed R to obtain streamflow at the river mouths.

2.1 The Canadian Regional Climate Model (CanRCM4)

CanRCM4 uses the fourth-generation Canadian atmospheric physics (CanAM4) package (von Salzen et al., 2013), which is the product of a multi-decadal program of climate model development at the Canadian Centre for Climate Modelling and Analysis (CCCma), a section within Environment and Climate Change Canada. The CanAM4 atmospheric physics package is also used in CanESM2 (Arora et al., 2011) which contributed results to the CMIP5. The difference between CanRCM4 and CanESM2, other than the former being a regional climate model and the latter being a comprehensive global ESM, is that CanRCM4 employs the limited-area

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configuration of the Global Environmental Multiscale (GEM) model (Côté et al., 1998), which uses a semi-Lagrangian dynamical core for advection in the atmosphere and is developed by Environment and Climate Change Canada's Recherche en Prévision Numérique (RPN) where it is used both for global and regional numerical weather prediction. CanESM2 on the other hand uses a spectral dynamical core for advection in the atmosphere. CanRCM4 is driven at its boundaries with data from its parent model (CanESM2). An overview and technical details of the coordinated global and regional climate modelling effort used to develop the CanESM2-CanRCM4 system are described in detail by Scinocca et al. (2016). Results from the model's North American 0.22° domain, for a single ensemble member, are primarily used here. In addition, we also used runoff from CanRCM4 0.44° resolution simulations for the North American domain because of the availability of a large ensemble (LE) of 50 members (CanRCM4 LE) (ECCC, 2018). The large ensemble simulations allow the consideration of CanRCM4's internal variability, which is an intrinsic property of the climate system and models, that is largely irreducible and could account for a large fraction of the inter-GCM model spread (Deser et al., 2020). The results used here from CanRCM4's form part of its contribution to the coordinated regional climate downscaling experiment (CORDEX) effort. The North American domain of CanRCM uses a rotated latitude-longitude projection with the North Pole at longitude 83° E and latitude 42.5° N, as opposed to the geographic North Pole (latitude 90° N).

The land surface component in the physics component of CanAM4 is the coupled CLASS-CTEM model. The physical processes are based on the Canadian Land Surface Scheme (CLASS) (Verseghy, 1991; Verseghy et al., 1993), and biogeochemical processes (which simulate vegetation as a dynamic component of the climate system) are based on the Canadian Terrestrial

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Ecosystem Model (CTEM) (Arora and Boer, 2005, 2003). The configuration of CLASS-CTEM used in CanESM2 and CanRCM4 uses three soil layers with thicknesses of 0.10, 0.25, and 3.75 m. Liquid and frozen soil moisture contents, and soil temperature, are determined prognostically for the three soil layers. The temperature, albedo, mass, and density of a single-layer snow pack (when environmental conditions permit snow to exist) are also prognostically modelled. Surface runoff is generated in CLASS when precipitation intensity exceeds infiltration capacity and when the top soil layer is saturated. The rainwater and snow melt that infiltrates the soil is available for soil evaporation and transpiration. Any remaining water percolates down the soil profile and comes out at the bottom of the soil profile and is termed drainage. Combined surface runoff and drainage constitute total runoff. Like most land surface components of ESMs, CLASS does not include a groundwater representation. Surface runoff and drainage components of runoff from CLASS are used as input into a large-scale river routing scheme to route runoff and obtain streamflow at the mouth of the rivers considered in this study as explained in the next section.

2.2 Variable velocity routing model

The variable velocity river routing scheme of Arora and Boer (1999) that is implemented in the family of Canadian ESMs (CanESMs) (Arora et al., 2009, 2011; Swart et al., 2019) is used to route daily runoff from CanRCM4. This routing scheme has been implemented in various versions of CanESMs at a spatial resolution of 2.81° since the year 2000. For this study, the routing scheme was implemented at a spatial resolution of 0.5°. The reason for using river routing at 0.5° resolution instead of scaling river networks to the 0.22° rotated latitude-longitude projection of CanRCM4 ‘s North American domain is that scaling river networks is a non-trivial task that cannot be fully automated (Arora and Harrison, 2007). In contrast, conservatively regridding runoff from

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one spatial resolution to another is a straightforward process. The routing scheme needs river flow directions and these are obtained from the Total Integrating Runoff Pathways (TRIP) dataset (http://hydro.iis.u-tokyo.ac.jp/~taikan/TRIPDATA/TRIPDATA.html, last accessed July 2023) of Oki and Sud (1998). The TRIP data are available at the regular latitude-longitude grid with the geographic North Pole at its usual location (90° N). Figure 1 shows the river networks at 0.5° resolution based on TRIP data which also identifies the six river basins investigated in this study. The Fraser River (identified by the light green colour) appears to have a river mouth over land. This is because the Fraser River drains into the narrow Strait of Georgia which is not resolved at the 0.5° resolution of the TRIP dataset. In addition, the TRIP data set does not resolve any inland lakes and provides river flow directions over grid cells that are lakes. This is in fact helpful because it avoids discontinuities in the river network.

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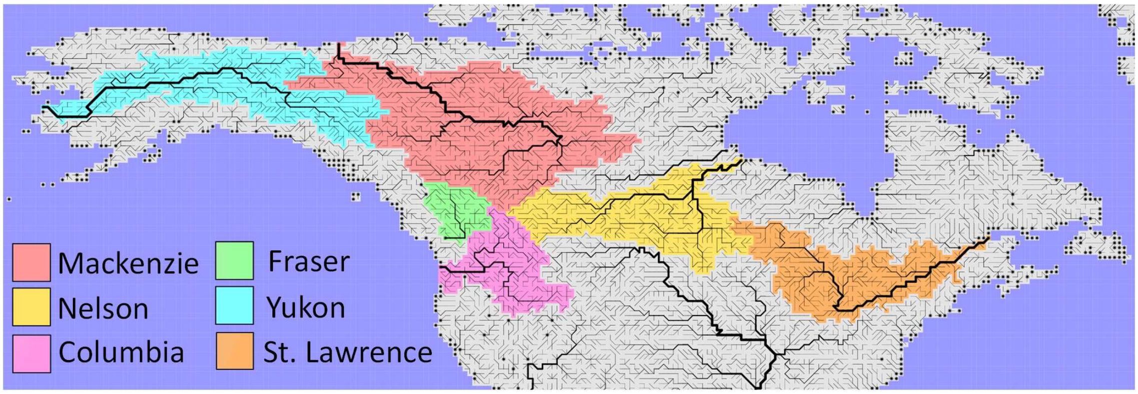

Figure 1: River flow networks at 0.5° resolution used in this study. The major river basins for which streamflow and runoff are analyzed in this study are also identified.

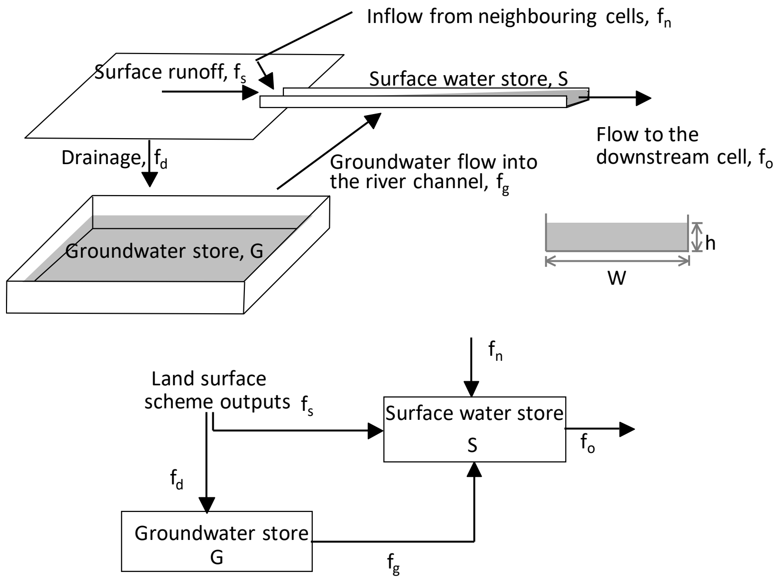

Figure 2: Schematic of the Arora and Boer (1999) river routing scheme used in this study to route runoff simulated by CanRCM4.

Figure 2 shows the schematic of the routing scheme which uses surface runoff and drainage outputs from the land surface scheme. The routing scheme is described briefly here and

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more details can be found in Arora and Boer (1999). The river channel is assumed to be rectangular and the width (W) of the river at every point along the river network is specified a priori. This river width is calculated based on its geomorphological relationship with mean annual discharge. The surface runoff contributes directly to the surface water store which is essentially the amount of water in the rectangular river channel between two grid cells. The flow velocity (*V*) is calculated using the Mannings formula.

$$V = \frac{1}{r} R^{2/3} s^{1/2} = \frac{1}{r}\left(\frac{A}{P}\right)^{2/3} s^{1/2} = \frac{1}{r}\left(\frac{Wh}{W+2h}\right)^{2/3} s^{1/2} \qquad (2)$$

where *r* is the Mannings roughness coefficient (a default value of 0.04 is used), *A* is the area of the river channel, *P* is the wetted perimeter, and *h* is the depth of water in the channel. Multiplied with the river's cross-sectional area, the time-varying velocity determines the output discharge from the surface water store of the current grid cell to the river channel of the downstream grid cell. The drainage from the bottommost soil layer contributes to the groundwater store which eventually contributes to the surface water store in the same grid cell. The delay in the groundwater store is based on the dominant soil texture type and is set to 10, 35, and 65 days if the dominant soil type in each grid cell is sand, silt, and clay, respectively, following Arora and Boer (1999). Both the depth and velocity of the water in the river channel are prognostic variables and evolve in time depending on the amount of water in the river channel.

The routing scheme used here does not consider the flow regulation effect of dams and reservoirs. It, however, does consider the effect of lakes and ice jams in a simple manner. The global lake data set from Kourzeneva et al. (2012) is used which prescribes the fractional coverage of sub-grid lakes and the five Laurentian Great Lakes (Lakes Superior, Michigan, Huron, Ontario,

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and Erie). In particular, the flow at the mouth of the St. Lawrence River is affected significantly by the Great Lakes. The hydraulic residence of water in the Great Lakes varies from about 2 years for Lake Erie to about 200 years for Lake Superior (Quinn, 1992). As a result, even in the absence of anthropogenic flow regulation for the St. Lawrence River, we expect the streamflow at its mouth to show very little seasonality compared to the usual spring peak of Canadian rivers dominated by snowmelt. The simple approach used here delays the streamflow flowing into a grid cell with a lake fraction greater than 60% using an e-folding time scale of 300 days similar to the treatment of the groundwater reservoir (Figure 2) (Arora and Boer, 1999). For the St. Lawrence River, the effect of delay caused by the Great Lakes is much larger than that of the anthropogenic flow regulation.

Ice jams and breakups are complex thermal and mechanical events and therefore challenging to model. They occur on all Canadian rivers with varying degrees and depend on winter temperatures, the river bathymetry, and the physical and geomorphological conditions of rivers (Prowse, 1986; Beltaos, 2000). The winter freezing of river water inevitably leads to a slow down of river flow velocity. When water cannot move downstream, upstream flooding results. Here, we have used a simple approach that increases Manning's roughness coefficient (a value of 0.08 is used) for the Mackenzie and the Yukon Rivers (which are the most northerly and therefore affected the most by ice jams) for the period January to June. Chen and She (2020) report the trend in river ice breakup dates for the Mackenzie and Yukon Rivers to be around -0.3 and -1.3 days/decade for the 1950-2016 period, where the negative sign indicates that the ice breakup is occurring earlier. Assuming the same trend, the breakup dates would occur about 2.5 and 11 days earlier towards the end of this century, respectively, for the Mackenzie and Yukon

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rivers. This simple approach reduces the river flow velocity during the months that are most affected by river ice jams. Although this is not a perfect nor a complete approach this simple treatment allows to improve the streamflow seasonality for the Mackenzie and Yukon rivers. For the southerly Fraser and Columbia rivers such treatment wasn't necessary. Consideration of a higher roughness coefficient for the St. Lawrence River to account for ice jams does not affect its streamflow's seasonality (or rather lack of it) which is overwhelmingly determined by the delay and storage caused by the Great Lakes.

2.3 Modelled and observation-based data

The CMIP5 historical simulation covers the period 1850-2005 and the future scenarios cover the period 2006-2100. We used daily runoff from CanRCM4 from its 0.22° North American domain for the 20 years 1986-2005 from one ensemble member of the historical simulation and for the 20 years 2081-2100 from one ensemble member each for the two future scenarios (RCP 4.5 and RCP 8.5, Moss et al. (2010)). The RCP 8.5 is the highest baseline emissions scenario where future development is based on continuous fossil-fuel development. As a result, CO_2 emissions and concentrations increase throughout the 21st century and CO_2 concentration in the year 2100 is around 1100 ppm. RCP 4.5 is a moderate emissions scenario in which emissions peak around 2040 and then decline: as a result CO_2 somewhat stabilizes to around 550 ppm by the year 2100. Since the CanRCM4 data are available on a rotated latitude-longitude grid and the river routing is performed on a regular latitude-longitude grid (following the TRIP data) the runoff data from CanRCM4 are conservatively regridded to the global 0.5° grid using climate data operators (CDO) (https://code.mpimet.mpg.de/projects/cdo/embedded/index.html#x1-7170002.12.5, last accessed Dec 2023). These runoff data are then used as input into the routing model. The 20-

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year runoff data (1986-2005 for the historical simulation, and 2081-2100 for the future scenarios) are concatenated into a 40-year time series for each simulation (historical, RCP 4.5, and RCP 8.5). These data are then input into the routing model and the last 20 years of simulated streamflow are then analyzed. The 20-year spin-up is sufficient to allow the surface and groundwater stores to fill up and reach equilibrium. The simulated precipitation and temperature from CanRCM4 are compared against observation-based data from the CRU TS 4.07 product (Harris et al., 2020).

The simulated streamflow is compared against observation-based estimates obtained from the Global Runoff Data Centre (GRDC) for the stations that are closest to the river mouths. Table 1 lists the drainage areas of all rivers considered in this study as discretized in the TRIP data set and at the stations closest to the river mouth. For the Columbia River, which is heavily regulated, we obtain an estimate of the naturalized flow with no regulation and no irrigation provided by the Bonville Power Administration (BPA) for the station VAN (near Vancouver, Washington, USA) (https://www.bpa.gov/energy-and-services/power/historical-streamflow-data;https://www.bpa.gov/-/media/Aep/power/historical-streamflow-reports/historic-streamflow-nrni-flows-1929-2008-corrected-04-2017.csv, last accessed July 2023). The drainage area of the Columbia River upstream of the VAN station is 616960 km^2 and does not include discharge contributions from three tributaries (Willamette, Cowlitz, and Lewis Rivers). Of these three tributaries, the contribution from Willamette is the largest. We obtained naturalized streamflow for the Willamette River at the station SVN (drainage area 25,600 km^2) also from BPA's website (https://www.bpa.gov/-/media/Aep/power/historical-streamflow-reports/correction-20220801.zip, from the file SVN6ARF_daily_COR.xlsx) and added it to the naturalized streamflow at the station VAN. This yields naturalized streamflow for the entire

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Columbia River basin, except the smaller Cowlitz, and Lewis Rivers, and represents a drainage area of 642,560 km^2 (see Table 1).

The Nelson River is affected by two large lakes, Lake Winnipeg and Lake Manitoba, and in addition, it is also heavily regulated. It currently has five dams towards the end of its journey as it flows into Hudson Bay. There are no upstream gauging stations close to the first upstream dam. In addition, water is also diverted from Churchill to the Nelson River. We were unable to obtain naturalized flow for the Nelson River. Due to anthropogenic flow regulation on the Nelson River, the present-day streamflow shows very little seasonality (as shown later). As a result, we do not evaluate the simulated monthly streamflow for the Nelson River and focus only on its mean annual value.

Table 1: Comparison of river basin areas as represented in the TRIP data and at the gauging station closest to the river mouth for the river basins considered in this study as obtained from the GRDC.

River basin	River basin area (million km^2)		Gauging station
	in the TRIP dataset	**at the gauging station closest to the river mouth**	
Mackenzie	1.74	1.66	Arctic red river
Yukon	0.85	0.83	Pilot Station
Columbia	0.66	0.64	See section 2.3
Fraser	0.23	0.22	Hope
Nelson	1.07	1.06	Long Spruce generating station
St. Lawrence	1.11	0.77	Cornwall, Ontario

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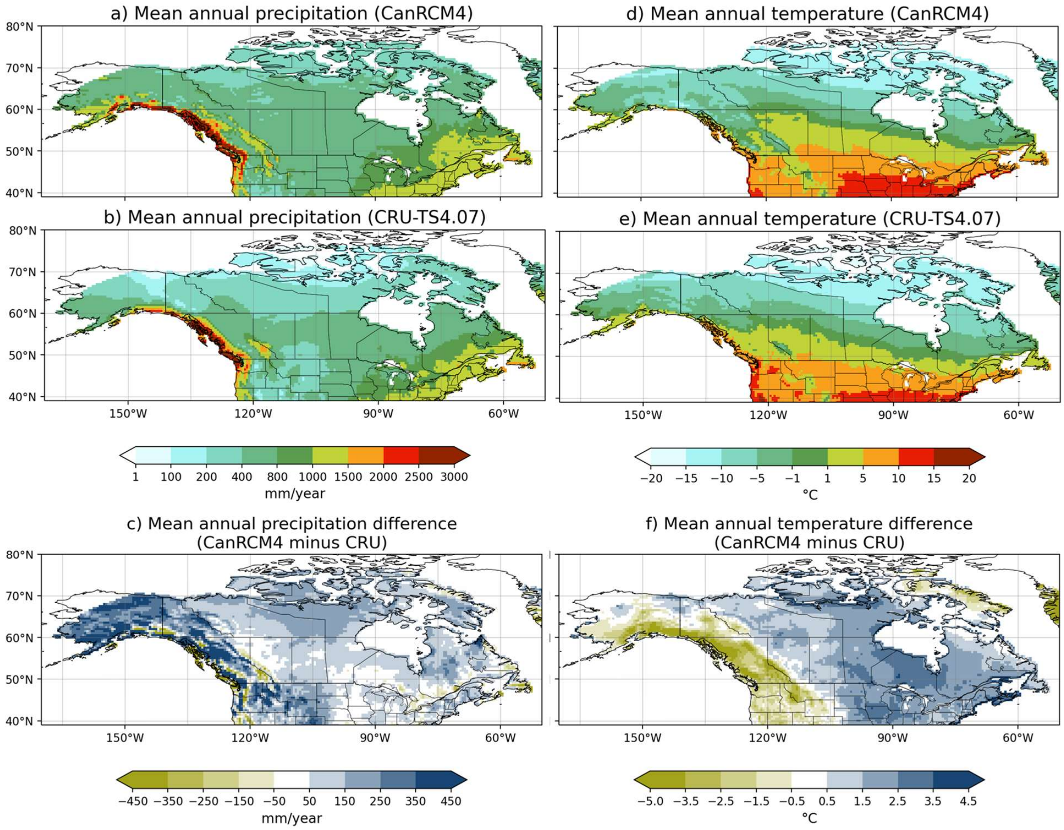

Figure 3: Comparison of CanRCM4 simulated precipitation (left column) and temperature (right column) with observation-based estimates from the CRU TS 4.07 dataset for the period 1986-2005.

3. Results

3.1 Present-day precipitation, temperature, and streamflow

Figure 3 compares the mean annual precipitation (left column) and temperature (right column) simulated by CanRCM4 to observation-based estimates from the CRU TS 4.07 dataset (referred to as CRU from here on) for the 1986-2005 period. Although the six river basins considered in this study do not cover the entire Canadian region, for completeness the plots are

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shown for the whole of Canada and south up to 39 °N to include the southern edge of the Columbia River basin. In Figure 3, while CanRCM4 broadly simulates the geographical distribution of temperature and precipitation reasonably realistically, there are differences compared to the CRU dataset. CanRCM4 generally simulates higher precipitation over Canada and more so to the west of the Rockies (Figure 3c) compared to observations. The model simulates cooler than observed temperatures to the west of the Rockies and higher than observed temperatures to the east of the Rockies (Figure 3f). This is likely related to the representation of topography in the model. The overall somewhat higher precipitation in CanRCM4 over North America is also noted by Alaya et al. (2019) who compared probable maximum precipitation (PMP) calculated using CanRCM4 data and compared it to estimates based on several reanalysis. Alaya et al. (2019) concluded that among the three reanalyses they considered, CanRCM4 compared best with the National Centre for Environmental Prediction's (NCEP) Climate Forecast System Reanalysis.

Figure 4 compares the simulated annual cycle of temperature (left column) and precipitation (middle column) over the six river basins (Figure 1) selected in this study with observation-based estimates from CRU. The right-hand side column compares simulated streamflow for the six river basins with observation-based estimates from the GRDC. The basin-averaged values of temperature and precipitation are calculated by area weighting the values in the individual grid cells that lie inside a given river basin according to the TRIP data (Figure 1). The plots also show the mean annual values (dashed lines) on the plot and their magnitude in the legend. Figure 4 shows that overall CanRCM4 simulated basin-wide averaged temperatures compare reasonably well with observation-based estimates based on the CRU for the Mackenzie and the Yukon River basins. For the Columbia and Fraser, the simulated temperatures are lower

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for most months, and for the Nelson River basin, the CanRCM4 simulated temperatures are higher compared to the CRU data. The seasonal cycle of temperature compares well with the observation-based estimates from CRU data. Compared to temperature, there are larger differences in simulated CanRCM4 precipitation compared to the CRU data. Although CanRCM4 simulates the seasonality of precipitation reasonably well compared to the CRU data, simulated precipitation is higher for all river basins, consistent with Figure 3c. The comparison with the CRU data provides useful insights into simulated quantities. However, all observation-based data sets (including CRU) have their limitations. Wong et al. (2017) compared several gridded observation-based precipitation datasets over Canada and found that they all have limitations and the datasets compared best with gause-based precipitation data in summer, followed by autumn, spring, and winter in order of decreasing quality. Sun et al. (2018) compare global precipitation from 22 gauge-, satellite-, and reanalysis-based products, including CRU, and quantify the uncertainty in the different precipitation estimates over timescales ranging from daily to annual. Shi et al. (2017) evaluated the CRU precipitation over large regions of China and found that CRU underestimates precipitation in that region compared to rain gauge records. In the end, the objective of the comparison of the simulated climate with CRU observations is to evaluate if the model climate is reasonably realistic for the present day. The assumption behind using direct output from climate models is that despite the biases in the simulated current climate it is possible to deduce meaningful information about the effect of climate change using the change in simulated quantities.

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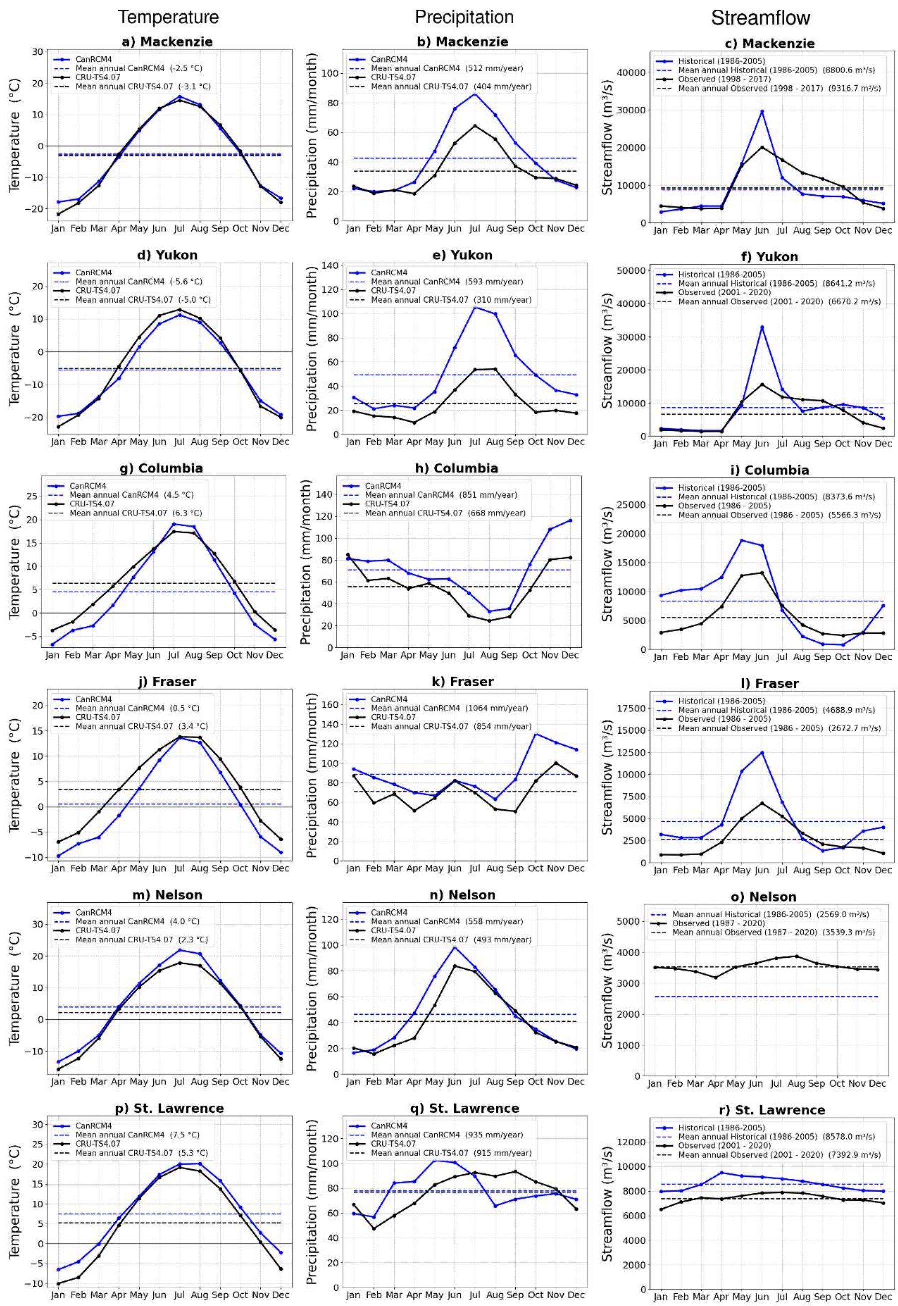

368 **Figure 4**: Comparison of the annual cycle of basin-wide averaged CanRCM4 simulated
369 temperature (left column) and precipitation (middle column) with observation-based estimates
370 from the CRU TS 4.07 dataset for the period 1986-2005. The right-hand side column compares
371 simulated streamflow with observations from the GRDC.

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The differences in simulated climate for the present day affect simulated streamflow as expected. The simulated mean annual streamflow is higher for four out of six river basins considered (Yukon, Columbia, Fraser, and St. Lawrence) primarily because of the higher simulated precipitation. Simulated precipitation is also higher for the Mackenzie River basin, but the mean annual simulated streamflow compares well with its observation-based estimate. Possible reasons for reasonably realistic annual simulated streamflow despite higher precipitation could be biases in the CRU dataset itself, or higher simulated evaporation in CanRCM4 (although simulated summer temperatures compare well with the CRU data). Finally, the simulated mean annual streamflow for the Nelson River is lower than its observation-based estimate despite somewhat higher simulated precipitation than the CRU data. The most likely reason for this is the diversion from the Churchill River into the Nelson River which started in 1976 to increase the water flow to larger generating stations on the lower Nelson River. The Manitoba government estimates that an average of 25% more water flows into the lower Nelson River due to the Churchill River Diversion (CRD) (https://www.gov.mb.ca/sd/water/water-power/churchill/index.html, last accessed Sep. 2023). The seasonality of Mackenzie, Yukon, and Fraser Rivers is dominated by the spring snowmelt with the peak occurring in June for both simulated and observed streamflow. The simulated streamflow for the Columbia and Fraser rivers peaks at the right time but there is more simulated streamflow during the winter months when precipitation is also higher than observed. For the Mackenzie and Yukon rivers although the mean annual simulated and observed streamflow are comparable their seasonal distribution is not. The simulated streamflow peak for these rivers is higher due to the simple treatment of ice jams which is not sufficient to hold the water in the river channel and then release it slowly

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as ice jams slowly dissipate in the spring and summer months, as the observed streamflow indicates. Finally, for the St. Lawrence River, there is little seasonality in observed streamflow due to the delay caused by the Great Lakes and anthropogenic flow regulation. The lack of strong seasonality simulated in simulated streamflow for the St. Lawrence River is caused entirely due to the delay caused by the Great Lakes (section 2.2).

Overall the spatial distribution of precipitation and temperature over Canada (Figure 3), and the seasonality of these two primary climate drivers for the river basins considered in this study (Figure 4), compare reasonably well with observation-based estimates from the CRU data, although there are differences in the absolute magnitude of these variables. The resulting seasonality of streamflow has limitations due to three factors: 1) the biases in the driving climate from CanRCM4, 2) the biases in the land surface component of CanRCM4 which partitions precipitation into evaporation and runoff, 3) the lack of calibration of the land surface component to specific river basins, and 4) the lack of processes in the routing component including the limitation of not being able to treat ice jams comprehensively. Despite these limitations, the simulated streamflow captures the broad seasonal patterns with higher values during the spring snow melt and lower values during the winter months as observations show.

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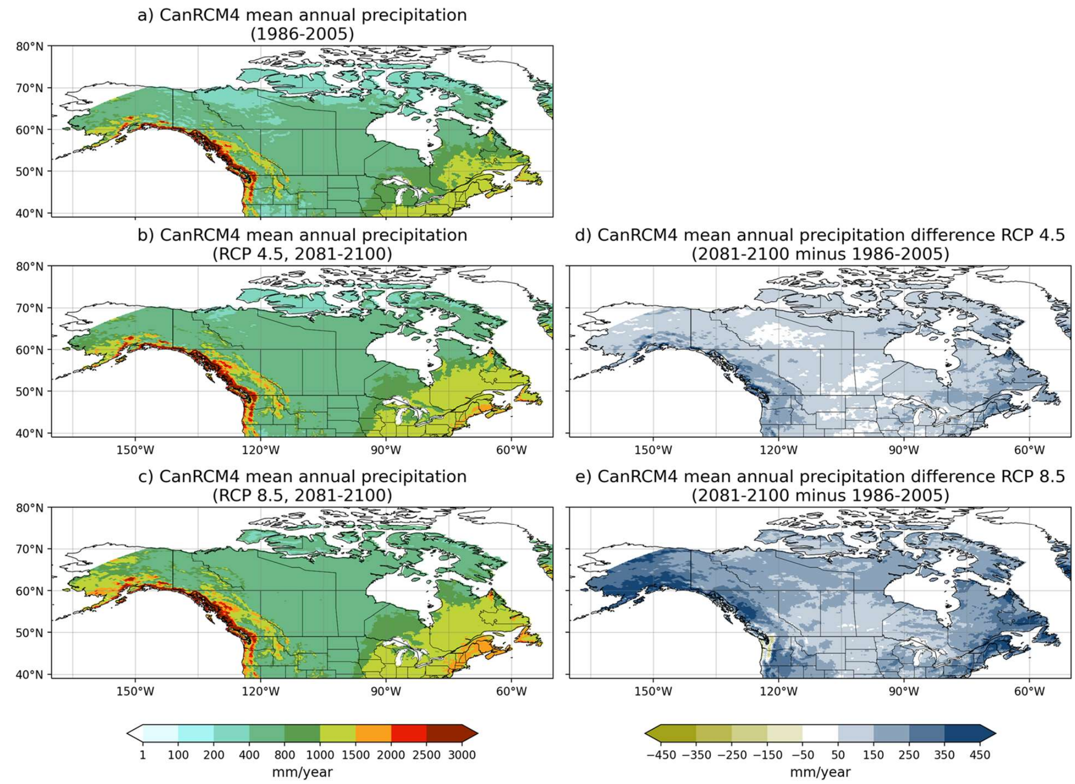

Figure 5: Comparison of CanRCM4 simulated precipitation for the 1986-2005 and 2081-2100 periods, for RCP 4.5 and 8.5 scenarios.

3.2 Changes in future climate and streamflow

Figures 5, 6, and 7 show the changes in CanRCM4 simulated precipitation, temperature, and runoff for the period 2081-2100, for both RCP 4.5 and 8.5 scenarios, compared to the 1986-2005 period from the historical simulation. Over Canada, simulated precipitation and temperature increase almost everywhere and in both scenarios. As expected, the magnitude of precipitation and temperature change is higher for the RCP 8.5 than the RCP 4.5 scenario. Simulated precipitation increases are higher in the coastal western and eastern Canadian regions than in central and northern parts of Canada. The central Canadian region sees the lowest

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increase in precipitation in both scenarios. Simulated temperature increases, as expected, are higher at higher latitudes due to polar amplification of the temperature change associated with the snow- and ice-albedo feedback. In the RCP 4.5 and 8.5 scenarios, the simulated temperature changes vary from about 3 °C and 6 °C, respectively, in the south, to about 6 °C and 11 °C, in the north. The parent climate model (CanESM2) on which CanRCM4 is based has an equilibrium climate sensitivity of 3.7 °C, somewhat on the higher side, compared to the range of 1.5 °C to 4.5 °C amongst climate models that contributed to CMIP5 (Schlund et al., 2020). As a result, we also then expect the magnitude of simulated changes to be somewhat higher than a model with average climate sensitivity.

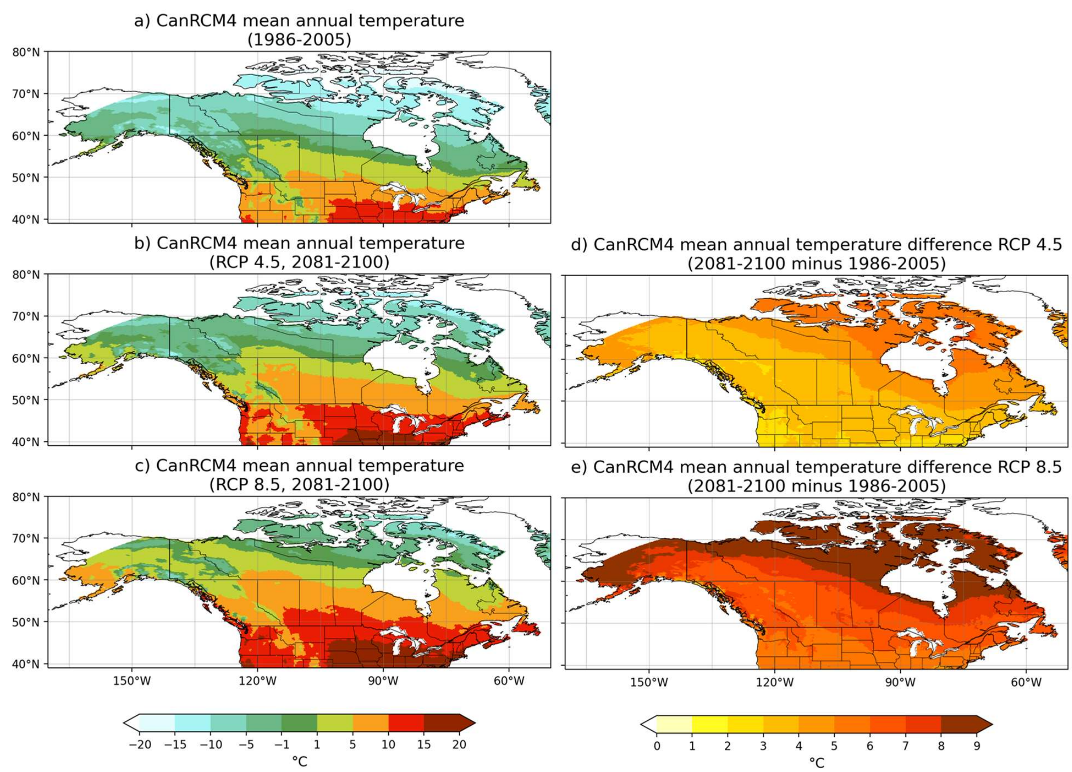

Figure 6: Comparison of CanRCM4 simulated temperature for the 1986-2005 and 2081-2100 periods, for RCP 4.5 and 8.5 scenarios.

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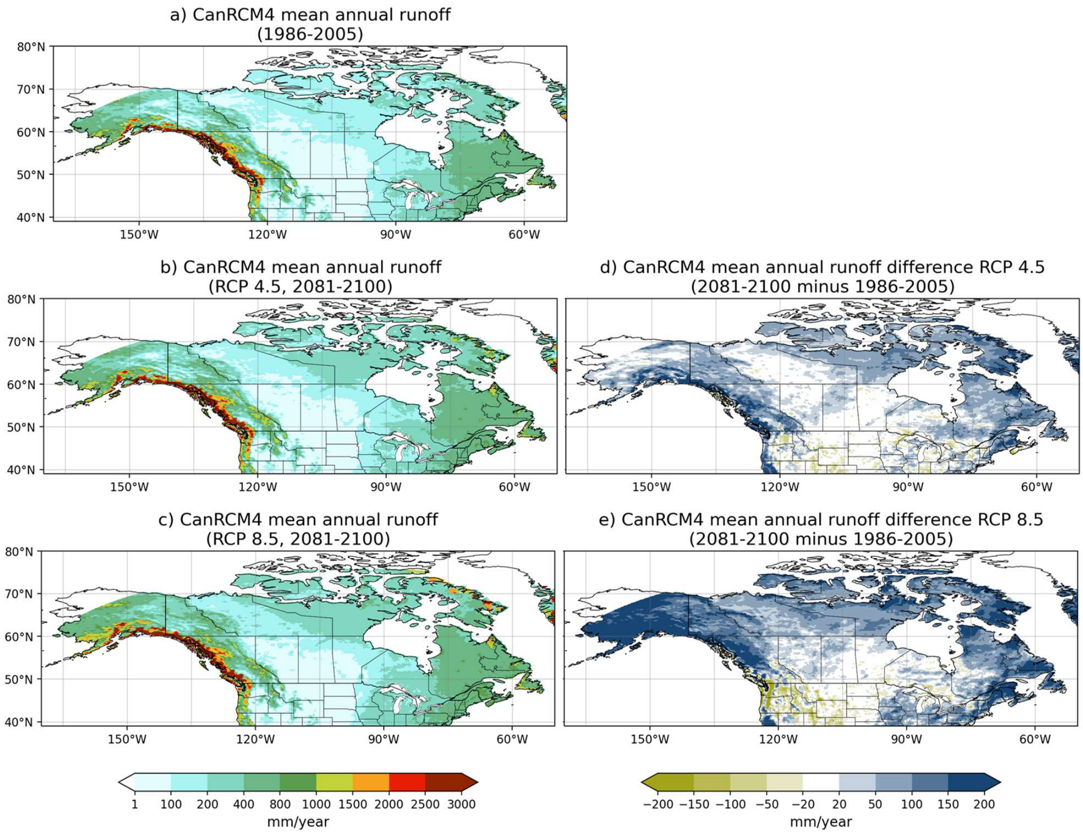

Figure 7: Comparison of CanRCM4 simulated runoff for the 1986-2005 and 2081-2100 periods, for RCP 4.5 and 8.5 scenarios.

In Figure 7 runoff increases generally everywhere in Canada for the RCP 4.5 and RCP 8.5 scenarios with larger changes on the west and east coasts, and in northern Canada, following a similar pattern of changes in precipitation. Runoff reduces in parts of the southern Columbia River basin in the United States in the RCP 4.5 scenario, and these decreases become more pronounced and widespread over the north-western Pacific region in the RCP 8.5 scenario including the Fraser River basin in Canada.

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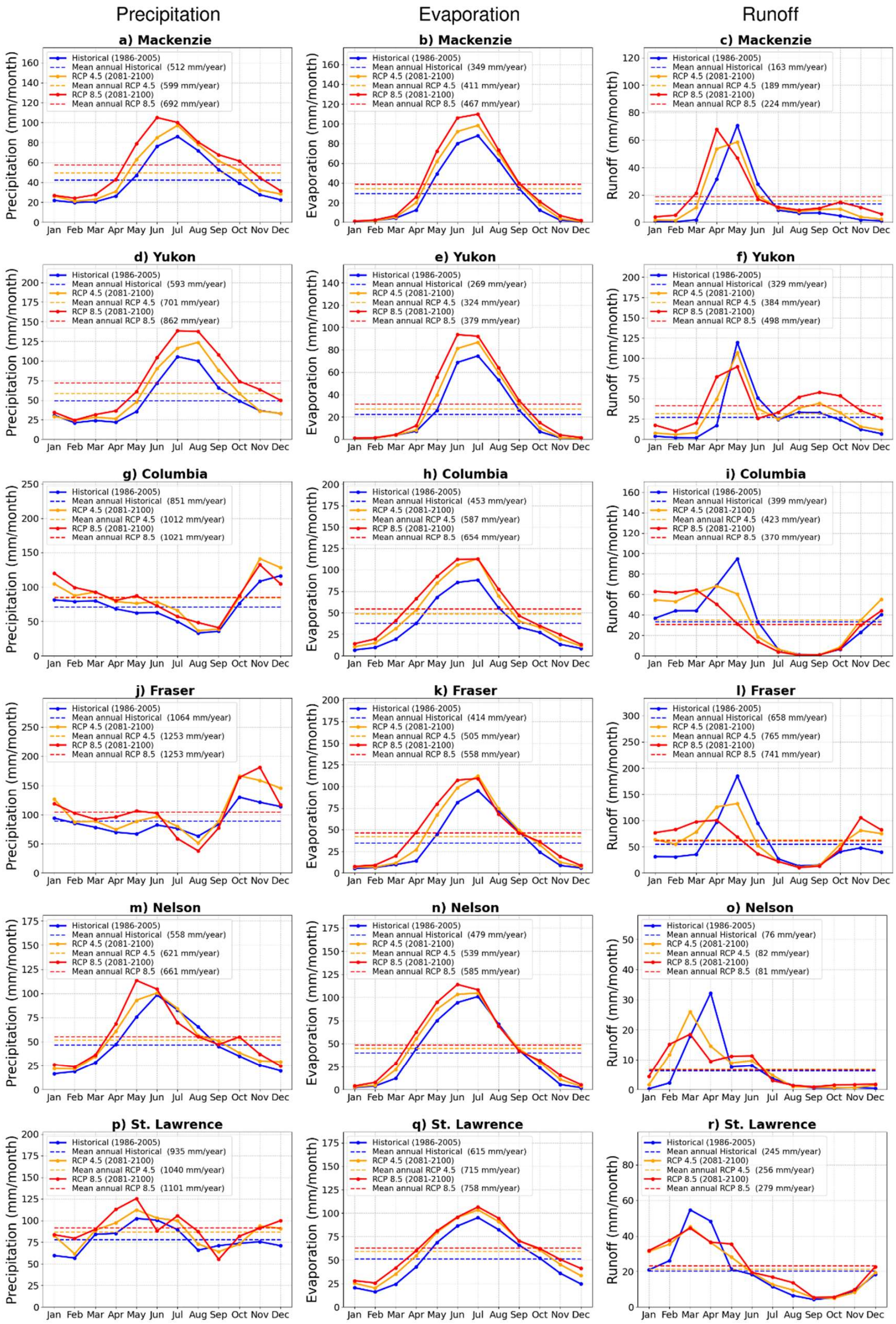

Figure 8: Comparison of the annual cycle of basin-wide averaged CanRCM4 simulated water budget components for each river basin for the historical (1986-2005) period and the two future scenarios RCP 4.5 and 8.5 (2081-2100): precipitation (left column), evaporation (middle column), and runoff (right column).

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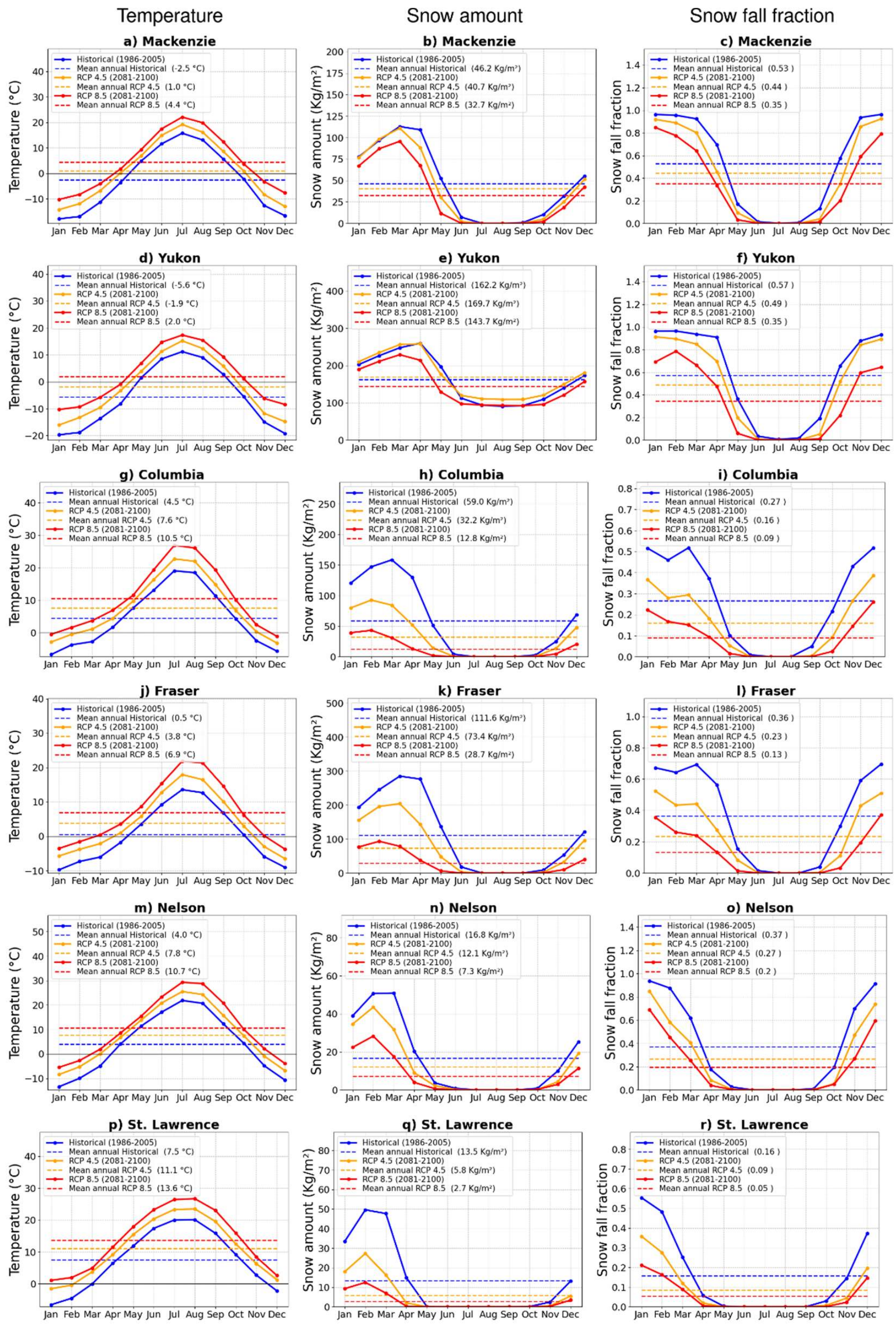

Figure 9: Comparison of the annual cycle of basin-wide averaged CanRCM4 simulated temperature (left column), snow water equivalent amount (middle column), and snowfall fraction (right column) for the historical (1986-2005) period and the two future scenarios RCP 4.5 and 8.5 (2081-2100).

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Figure 8 shows the annual cycle of the simulated water budget components (precipitation, evaporation, and runoff) for the six river basins considered in this study for the historical (1986-2005) period and the two future scenarios, RCP 4.5 and 8.5 (2081-2100). As in Figure 4, the mean annual values are shown as dashed lines and their magnitude is noted in the legend.

Table 2: Evaporation and runoff ratios for the five river basins simulated by CanRCM4 for the historical period (1986-2005) and the two future scenarios (RCP 4.5 and 8.5, 2081-2100). The evaporation (runoff) ratio is the ratio of mean annual evaporation (runoff) to precipitation.

River basin	**Evaporation ratio (E/P)**			**Runoff ratio (R/P)**		
	Historical (1986-2005)	**RCP 4.5 (2081-2100)**	**RCP 8.5 (2081-2100)**	**Historical (1986-2005)**	**RCP 4.5 (2081-2100)**	**RCP 8.5 (2081-2100)**
Mackenzie	0.682	0.686	0.675	0.318	0.316	0.324
Yukon	0.454	0.462	0.440	0.555	0.548	0.579
Columbia	0.532	0.580	0.641	0.469	0.418	0.362
Fraser	0.389	0.403	0.445	0.618	0.611	0.591
Nelson	0.858	0.868	0.885	0.136	0.132	0.123
St. Lawrence	0.664	0.686	0.684	0.314	0.294	0.302

The evaporation (E/P) and runoff (R/P) ratios for the six river basins for the historical period and the two future scenarios are shown in Table 2 and allow to see how the partitioning of precipitation into evaporation and runoff changes with climate. For the mean annual values of P, E, and R reported in Figure 8, P is balanced to within 1% by (E+R) for all river basins (except the St. Lawrence) and all scenarios, except for the Yukon (for RCP 8.5) and the Fraser River basins (for RCP 4.5 and 8.5) for which (E+R) is higher than P indicating that ΔS is not zero (see equation 1). As a result, (E/P) and (R/P) also add to one for all river basins except for the Yukon (RCP 8.5,$(E+R)/P = 1.02$) and the Fraser River (RCP 4.5, $(E+R)/P = 1.014$, and RCP 8.5, $(E+R)/P = 1.036$) basins. For the St. Lawrence River basin, the imbalance is around 2% because of the presence of the Great Lakes which had to be excluded from the river basin mask.

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Since basin-wide averaged calculations are done at 0.5° latitude-longitude resolution, and the actual domain of CanRCM4 is on rotated latitude-longitude projection this led to slightly more rounding errors for the St. Lawrence than other river basins.

For all river basins considered, precipitation increases for both future scenarios with the increase being larger for the RCP 8.5 scenario consistent with Figures 5d and 5e. The response of evaporation to changes in climate is expected. The increase in precipitation and temperature yields an increase in evaporation for future scenarios for all river basins. Simulated runoff does not increase as much as precipitation since evaporation also increases. The runoff ratio, in Table 2, increases for the northerly Mackenzie and the Yukon River basins while it decreases for the southerly Nelson, St. Lawrence, and especially for the Fraser and Columbia River basins which are characterized by milder climate owing to their location in the Pacific north-western region. This is because the increase in precipitation is more than enough to compensate for the increase in evaporation (associated with a warmer climate) for the northern river basins but not for the southern ones (as seen earlier in Figure 7 where runoff begins to decrease in parts of the Columbia and Fraser River basins). The absolute runoff amount in Figure 8 increases for the Mackenzie and Yukon River basins, in the RCP 4.5 and 8.5 scenarios compared to the historical simulation, but doesn't change much for the Columbia, Fraser, Nelson, and St. Lawrence River basins. However, the seasonality of runoff changes for all river basins, and simulated runoff peak either occurs earlier in the year, occurs with reduced magnitude, or both. Canadian rivers are dominated by spring snowmelt and this runoff behaviour is associated with snow melt occurring earlier in the year in the RCP 4.5 scenario than in the historical simulation, and occurring earlier in the RCP 8.5 scenario than in the RCP 4.5 scenario. This is seen in Figure 9 which shows the

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simulated annual cycle of temperature changes, snow amount, and snowfall as a fraction of total precipitation for the historical period and the two RCP scenarios for the six river basins. In Figure 9 the mean annual temperature increase from the historical period to the RCP 4.5 scenario, and from the RCP 4.5 to RCP 8.5 scenario, is between 3 and 3.5 °C for the six river basins considered here. The middle column of Figure 9 shows that in addition to earlier snowmelt the amount of snow in the winter months decreases for all river basins with climate warming. The only exception to this is the Yukon River basin in which the mean annual snow amount increases marginally in the RCP 4.5 scenario (Figure 9e). As expected, the fraction of precipitation falling as snow also decreases with climate warming for all river basins (right column, Figure 9).

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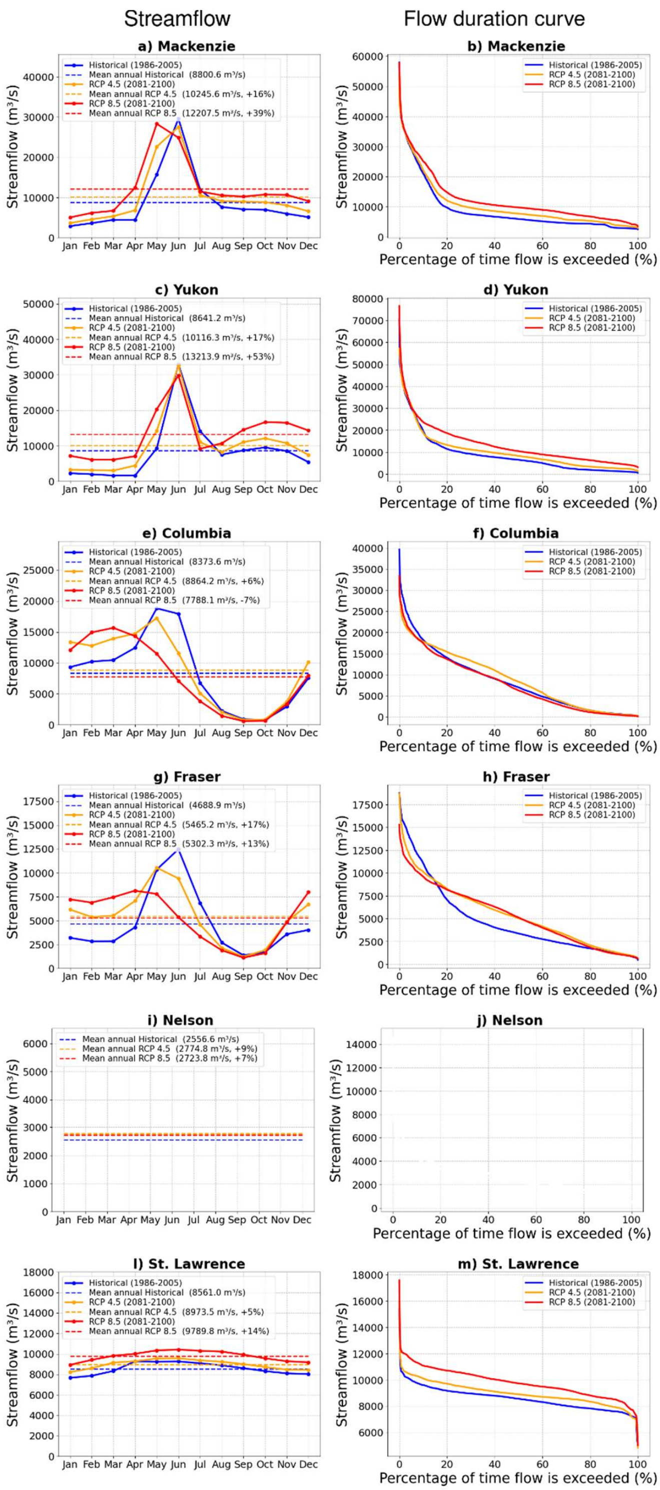

Figure 10: Comparison of the simulated monthly streamflow (left column) and flow duration curves (right column) for the historical (1986-2005) period and the two future scenarios RCP 4.5 and 8.5 (2081-2100) for the six river basins considered.

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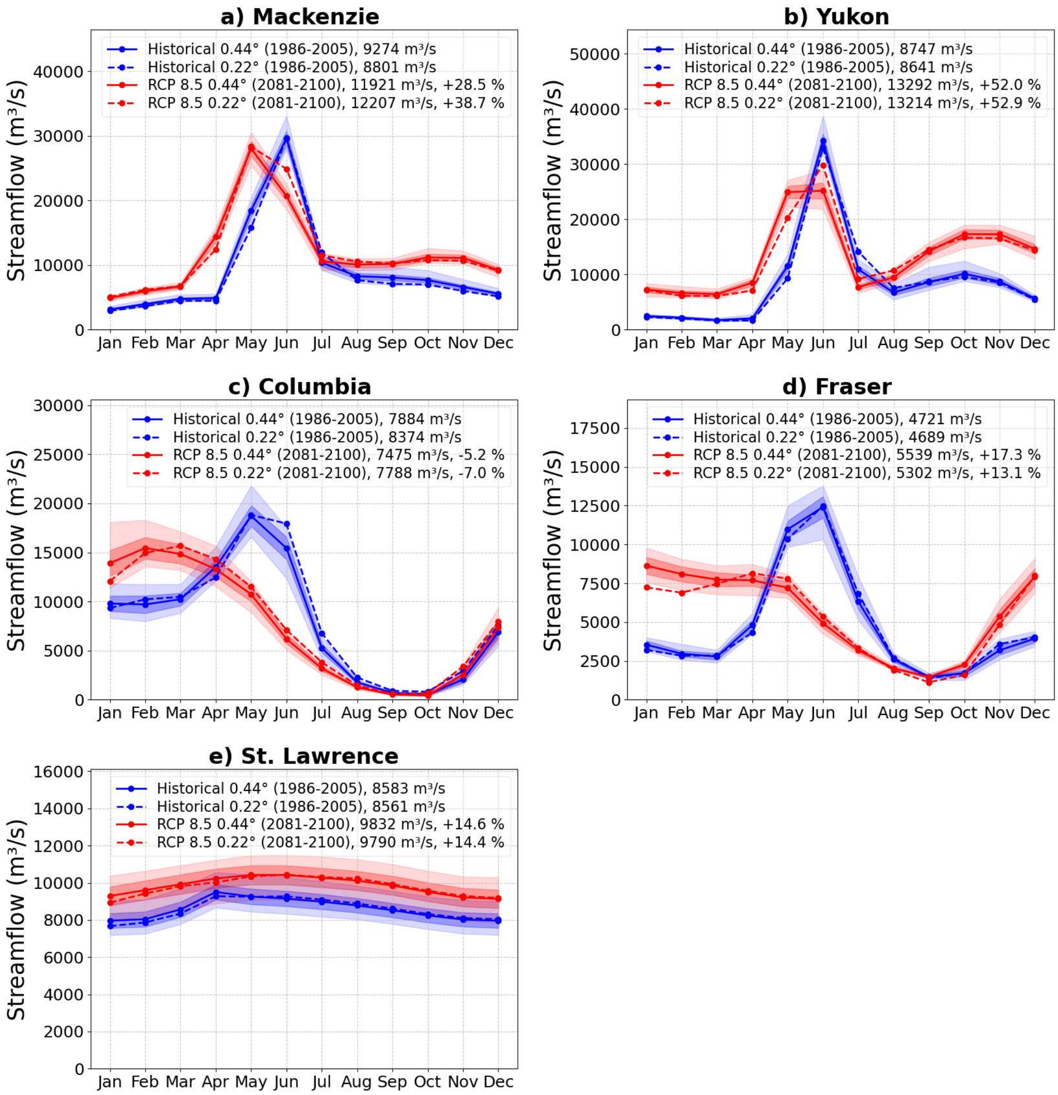

Figure 11: Comparison of the simulated monthly streamflow for the historical (1986-2005) period and the RCP 8.5 scenario (2081-2100) for the river basins considered in this study from the 0.22° and 0.44° simulations. The results from the 0.22° simulations (shown earlier in Figure 10) are shown as dashed lines. The uncertainty range for the 0.44° simulations is based on results from CanRCM4's 50-member large ensemble. The solid lines indicate the mean across 50 members the light shading indicates the full range, and the dark shading indicates the mean ± one standard deviation range, for the 0.44° simulations.

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Figure 10 compares simulated monthly streamflow and flow duration curves for the historical (1986-2005) period with those from the two future scenarios RCP 4.5 and 8.5 (2081-2100) for the six river basins considered. The flow duration curves are calculated using daily streamflow values. Monthly streamflow and flow duration curves are not shown for the Nelson River because we do not consider anthropogenic flow regulation, as mentioned earlier. The legends in Figure 10 for the streamflow figures in the left column show mean annual values but also the change from the simulated historical values for the RCP 4.5 and 8.5 scenarios. The mean annual streamflow increases for all rivers for both the RCP 4.5 and 8.5 scenarios, except for the Columbia River for the RCP 8.5 scenario (-7%). The increase in simulated annual streamflow is largest for the Mackenzie (+16%, +39%) and Yukon Rivers (+17%, +53%) for the RCP 4.5 and 8.5 scenarios, due to higher precipitation increase in these two basins (Figure 8). The increase in annual streamflow for other rivers is smaller and between 6% and 14%.

The changes in streamflow seasonality are larger for the southerly Columbia and Fraser Rivers than for the northerly Mackenzie and Yukon Rivers. The peak monthly streamflow for the Yukon River still occurs in June given it's the coldest river basin (Figure 4d) and the streamflow seasonality is still dominated by the spring snowmelt. However, despite the peak streamflow still occurring in June the streamflow does begin to increase earlier due to earlier snowmelt (Figure 9e). While the June streamflow peak doesn't change substantially, streamflow increases for most other months for the Yukon River. For the Mackenzie River, the peak streamflow occurs in June in the RCP 4.5 scenario as in the historical simulation but a month earlier in the RCP 8.5 scenario. Like the Yukon, although the peak streamflow doesn't change substantially for the Mackenzie River it increases for most other months. These changes in streamflow are also seen in the flow

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duration curves which show that for these two rivers the frequency of the occurrence of flows that occur greater than about 5% of the time in the historical simulation increases in the future. The Columbia and the Fraser Rivers experience much larger changes in their seasonality as their primarily snow-dominated nival flow regimes change to more hybrid flow regimes. The snowmelt-driven streamflow peak in spring is reduced considerably for future scenarios since a lower fraction of fall, winter, and spring precipitation falls as snow. As a result, streamflow increases from October to April since precipitation falls as rain, as opposed to snow, yields runoff that runs straight into the rivers. Additionally, the large reduction in snowpack volume together with earlier melt (Figure 9k and 9h) affects the seasonality of Fraser and Columbia streamflow and causes pronounced shifts in peak flows. The pronounced changes in the Fraser River basin peak flow are apparent in its flow duration curve (Figure 10h) which shows a decrease (increase) in the frequency of streamflow events which occurred less (more) than about 16% of the time and result in a more equitable streamflow regime with a pronounced reduction in its seasonality. Simulated streamflow for the St. Lawrence River shows very little seasonality and since annual streamflow increases for both scenarios, the flow duration curve simply moves up (Figure 10m).

3.3 Uncertainty in simulated changes in future streamflow

In addition to the 0.22° simulations for the North American domain, simulation results are also available from the 50-member large ensemble (LE) of CanRCM4 at 0.44° resolution. The LE data are, however, only available for the historical simulation and the RCP 8.5 scenario. Similar to the 0.22° resolution, we regridded the 0.44° runoff at CanRCM4's rotated latitude longitude projection to 0.5° regular latitude longitude projection for use as input into the river routing

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scheme. The use of the results from the LE allows us to quantify the uncertainty associated with streamflow based on the spread in the simulated results associated with the internal variability of the CanRCM4 model. This is illustrated in Figure 11 which shows the simulated streamflow for all the rivers considered here except the Nelson River. In Figure 11, the solid lines show the average across the 50 members of the LE, light shading shows the full range of the results, and dark shading shows the mean ± one standard deviation range (this implies the 5%-95% range when assuming normally distributed monthly streamflow values). In addition, streamflow from the 0.22° simulations (from Figure 10) is shown as dashed lines to allow direct comparison of results from the 0.22° and 0.44° simulations.

The changes in simulated streamflow are consistent between the 0.22° and 0.44° simulations, and for the most part, the results from the 0.22° simulations lie within the full range of results from the 0.44° simulations. This is expected since the driving climate at the boundaries of CanRCM4 based on CanESM2 is the same in both resolutions. The exact values and the change from the historical to the RCP 8.5 scenario (see legend for individual rivers) are, however, somewhat different. This is also expected since these simulations are different from the 0.22° simulations. The coarser resolution of the 0.44° simulations implies that the topography is not as realistically represented as in the 0.22°simulations. In particular, for the Yukon, the results based on the 0.22° simulation indicate that the month and the magnitude of peak streamflow do not change significantly in the RCP 8.5 scenario (Figure 10c), while those based on 0.44° suggest that they do. There are some differences between 0.22° and 0.44° results for the Mackenzie and Fraser Rivers too. Overall, the LE from the 0.44° simulations helps to provide context for results from the 0.22° simulations.

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Overall, despite the differences in the magnitude of changes, the direction and variability of change obtained from this study is generally consistent with the previous studies using basin-scale hydrologic models, driven by statistically downscaled and bias-corrected climate model data, for instance for the Fraser River (Shrestha et al., 2012; Islam et al., 2019), the Columbia River (Schnorbus et al., 2014) and the Yukon River (Hay and McCabe, 2010). The results presented here are also comparable to the projections from global and regional scale hydrologic models, e.g. for the Mackenzie River basin (Krysanova et al., 2017, 2020).

4. Summary and conclusions

This study offers a consistent analysis of results across six river basins in Canada although based on results from a single climate model. Despite the biases in simulated present-day CanRCM4 climate, and some differences in the results based on 0.22° and 0.44° simulations, the results provide useful information about changes in simulated streamflow that is consistent with expectations of process behaviour in a warmer climate, and with published literature.

Neither future precipitation nor temperature changes are uniform across Canada. Simulated precipitation increases are higher closer to the west and east coasts, and simulated temperature changes are higher towards the Arctic. Similar to precipitation, runoff changes are also higher closer to the west and east coasts. The changes in simulated streamflow indicate how the present-day climate state of river basins plays a role in their response to climate change. The results yield two broadly distinct responses of monthly streamflow changes to climate warming, up until the end of this century, for the northerly Mackenzie and Yukon rivers and the southerly Fraser and Columbia rivers. Despite higher future projected temperature changes in Canada's

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north, peak streamflow for the Mackenzie and Yukon rivers is still dominated by the spring snowmelt. This is because the present-day colder states of these river basins imply that even after around 6-7 °C warming, the basin-wide average temperatures are cold enough to not sufficiently change their snowmelt-dominated streamflow regimes. Changes, however, do occur in streamflow seasonality for these two rivers. Peak streamflow month changes from June to May for the Mackenzie River in the RCP 8.5 scenario for both 0.22° (Figure 9a) and 0.44° (Figure 11a) simulations and for the Yukon River in the RCP 8.5 scenario for 0.44° simulations (Figure 11b). Other than the changes in peak month and peak streamflow, snowmelt occurs earlier so streamflow starts increasing earlier in the spring and the streamflow during the rest of the year also increases (except for July) due primarily to an increase in precipitation. In contrast, the streamflow seasonality for the southerly Fraser and Columbia rivers is significantly more affected by warmer temperatures because the mean annual basin-wide temperature for these river basins is already above 0° C for the historical period. Both these rivers experience pronounced changes in their streamflow seasonality. The peak streamflow for both rivers decreases considerably and occurs two months earlier for the Fraser River and about 2-3 months earlier for the Columbia River in the RCP 8.5 scenario. These results appear exaggerated compared to the 1-2 months earlier peak in previous studies for the Fraser (Shrestha et al., 2012; Islam et al., 2019) and Columbia (Schnorbus et al., 2014) rivers that used results from multiple climate models. Shrestha et al. (2021a) used CanRCM4 data to evaluate snowpack response to varying degrees of warming. They found that snowpack reduction using CanRCM4-LE is higher than the ensemble of results obtained by driving a hydrological model with data from other climate models (their supplementary information), consistent with CanESM2's higher climate sensitivity.

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The results presented here also appear to show that the simulated changes in streamflow are somewhat resolution-dependent. This would be expected especially for topography-dominated river basins. If a LE of 50 members for the 0.22° resolution was also available, it would have been easier to draw firm conclusions about the effect of the spatial resolution on changes in simulated streamflow.

There are two primary limitations of the work presented here. First, we use results from only one climate model. It would have been ideal to use runoff from other regional climate models to provide an uncertainty range based on the spread across different climate models. This would also allow us to evaluate how the spread across the models compares to the spread across the 50 members of the CanRCM4 LE. Second, the results are based on direct output from the CanRCM4 climate model and direct climate model output is biased. This limitation is tied to our methodology. Bias-correcting climate data in our case inevitably implies using a different hydrological model or land surface scheme, than the land surface component of CanRCM4, and forcing it with bias-corrected climate data to obtain runoff. Finally, there are uncertainties associated with the routing process itself. As mentioned earlier, the routing scheme accounts for ice jams in a simplified manner and anthropogenic flow regulation is not taken into account.

Large ensembles are now becoming more common. The challenge for similar future studies is to consider the inter-model and intra-model (based on ensemble members of the same model) spreads in the same framework to derive an uncertainty estimate that is consistent with both types of uncertainties.

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We thank Daniel Peters for the helpful discussions at the beginning of this work and Sal Curasi for providing comments on the final version of this manuscript. We also acknowledge the efforts of the climate modelling team at the Canadian Centre for Climate Modelling and Analysis (CCCma) who made the results from CanRCM4 available.

Code/Data availability

The CanRCM4 data from 0.22° simulations used in this study are available from CCCma website (https://climate-modelling.canada.ca/climatemodeldata/canrcm/CanRCM4/). The data from the 0.44° CanRCM4 large ensemble are available from Environment and Climate Change Canada (https://open.canada.ca/data/en/dataset/83aa1b18-6616-405e-9bce-af7ef8c2031c).

Author contributions

VKA designed the study and wrote the majority of the manuscript. AL implemented river routing to operate at 0.5° resolution and performed all the simulations. RS and AL contributed to the manuscript text. RS also performed a literature review of existing studies that focus on the impact of climate change on Canadian rivers.

Competing interests

The authors declare that they have no competing interests.

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