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Landscape and climate conditions influence the hydrological sensitivity to climate change in eastern Canada

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ABSTRACT

Hydrological conditions in cold regions have been shown to be sensitive to climate change. However, a detailed understanding of how regional climate and basin landscape conditions independently influence the current hydrology and its climate sensitivity is currently lacking. This study, therefore, compares the climate sensitivity of the hydrology of two basins with contrasted landscape and meteorological characteristics typical of eastern Canada: a forested boreal climate basin (Montmorency) versus an agricultural hemiboreal climate basin (Acadie). The physically based Cold Regions Hydrological Modelling (CRHM) platform was used to simulate the current and future hydrological processes. Both basin landscape and regional climate drove differences in hydrological sensitivities to climate change. Projected peak SWE were highly sensitive to warming, particularly for milder baseline climate conditions and moderately influenced by differences in landscape conditions. Landscape conditions mediated a wide range of differing hydrological processes and streamflow responses to climate change. The effective precipitation was more sensitive to warming in the forested basin than in the agricultural one, due to reductions in forest canopy interception losses with warming. Under present climate, precipitation and discharge were found to be more synchronized in the greater relief and slopes of the forested basin, whereas under climate change, they are more synchronized in the agricultural basin due to reduced infiltration and storage capacities. Flow through and over agricultural soils translated the increase in water availability under a warmer and wetter climate into higher peak discharges, whereas the porous forest soils dampened the response of peak discharge to increased available water. These findings help diagnose the mechanisms controlling hydrological response to climate change in cold regions forested and agricultural basins.

1. Introduction

Land cover can have distinctive influences on snow accumulation (Pomeroy and Gray, 1995). In boreal forests, up to 60% and 40% of cumulative snowfall can be intercepted and sublimated, respectively (Hedstrom and Pomeroy, 1998; Pomeroy et al., 1998). Compared to open areas, snowmelt rates can be up to 70% lower in forests because of attenuated incoming shortwave radiation and reduced sensible and latent heat fluxes resulting from dampened wind speed by canopies (Varhola et al., 2010). Within forests, while small clearings are sheltered by the nearby forest canopy, larger clearings can lose snow accumulation via blowing snow erosion, which can lead to less snow accumulation in clearings than in the adjacent coniferous forest (Broxton et al., 2015; Pomeroy et al., 2012; Pomeroy and Gray, 1995). Over open and wind exposed areas, blowing snow can transport and sublimate as much as 75% of the annual snowfall (Pomeroy and Gray, 1995).

Many cold regions have been reported to exhibit different sensitivities to temperature and precipitation change, depending on their current cold season temperature regime governed by latitude and/or elevation: more drastic changes in snow cover were found to occur over regions with near-freezing air temperatures, whereas colder regions were found to be comparatively less sensitive to climate change (Aygün et al., 2020a). In Northern Europe, for instance, more pronounced declines in snow accumulation are projected in response to warming in the coastal and southern regions as they have milder temperatures compared to the interior and northern regions (Arheimer et al., 2013; Kellomäki et al., 2010; Räisänen and Eklund, 2012; Stonevičius et al., 2017). Similarly, in North America, the most dramatic declines in snow

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accumulation are projected to occur in regions with mild cold seasons, such as the coastal regions of North America (Brown and Mote, 2009). Schnorbus et al. (2014) have found that while snow water equivalent (SWE) is projected to decline in the Peace, Campbell and Upper Columbia river basins in British Columbia, Canada, the reductions are projected to be most drastic in the lower elevation and warmer coastal Campbell watershed and less pronounced in the high elevation and colder interior Upper Columbia watershed. Köplin et al. (2012) related climate change signals to physiographic properties including mean altitude, slope, elevation range and dominant land use in 186 basins in Switzerland. They showed that hydrological changes are strongly correlated to altitude, which governs the mean annual temperature of a basin and the associated hydro-climatological processes, e.g., the ratio of snowfall to total precipitation and snowmelt. While these studies showed that the present-day climate of a region is an important indicator for explaining the responses of snow accumulation to climate change, they did not resolve how the dominant land cover of a basin will modify the response of snow accumulation and runoff to climate change.

Multiple studies have shown that forest harvesting can lead to increased peak flow as a result of higher snow accumulation due to reductions in snow interception by the canopy (Buttle et al., 2009; Moore and Wondzell, 2005; Pomeroy et al., 2012; Whitaker et al., 2002). Other studies also reported higher peak flows when the forest cover is reduced or converted to agricultural fields or urban areas, which was mostly explained by the increased surface runoff due to reduced infiltration capacities resulting from compaction of the soil (Brown et al., 2005; Chandler, 2006; Easton et al., 2007; Germer et al., 2010; Savary et al., 2009; Ziegler et al., 2004). Deforestation for agricultural or urban purposes has also been reported to be responsible for an increase in annual water yield (Brown et al., 2005; Dias et al., 2015; Savary et al., 2009), which is mainly due to reduced evapotranspiration (Robinet et al., 2018). These studies are helpful for understanding the impact of land cover changes on peak and annual flows, but the question as to how each component of the basin landscape, i.e. land cover, topography and soil conditions, affects its hydrological sensitivity to climate change has not been directly addressed by these studies.

Exploring the relationships between basin characteristics and flow signatures in Europe, Kuentz et al. (2017) have reported that land cover is one of the most important determining factors on controlling flow signatures. Padrón et al. (2017), on the other hand, have shown that vegetation related factors have only a minor role on the partitioning of precipitation into runoff and evapotranspiration, while climate-related variables are the key controls. While Williams et al. (2012) have found no evidence that regions with higher fractions of snowfall favor streamflow, Berghuijs et al. (2014) later demonstrated that a higher fraction of snowfall is associated with higher mean annual streamflow for basins throughout the contiguous US. Sankarasubramanian et al. (2001) have shown that the sensitivity of streamflow to changes in precipitation is lower in regions that have higher average annual snowpack depths compared to other regions in the US. Analyzing the climate sensitivities of floods across several British basins, Prudhomme et al. (2013) did not find any clear geographical patterns. On the contrary, Rice et al. (2015) found that the geographical locations of river basins are highly influential on the variability of streamflow trends across the continental US. They also showed that the topographic characteristics of basins (elevation and slope) are strongly related to streamflow trends within individual ecoregions. On the other hand, Sawicz et al. (2014) reported that land use provides no valuable information in describing the streamflow responses to climate change across the continental US. These contrasting conclusions drawn in the literature regarding the roles of climatic factors and basin properties on streamflow highlight the need for more studies on this topic.

A top-down approach, where data from climate models drive hydrological models, is often used to assess potential climate impacts on hydrology (Arnell and Gosling, 2016; Minville et al., 2008; Sulis et al., 2012; Teutschbein et al., 2015). However, hydrological modelling studies become computationally expensive due to the increasing number of climate models and must be repeated whenever new climate projections become available (Kay et al., 2021). Some, instead, have thus adopted a less time-consuming sensitivity-based approach, using existing climate change projections to calculate uniform seasonal or annual climatic changes over a region. This approach allows investigating a wider spectrum of simple climate change scenarios against which updated climate projections can be compared (Aygün et al., 2021; Aygün et al., 2020b; López-Moreno et al., 2013; Prudhomme et al., 2010; Rasouli et al., 2015; Rasouli et al., 2022; Wetterhall et al., 2011; Whateley et al., 2014).

The main objective of this study is to find out the degree to which regional climate and landscape (land cover, topography and soil conditions) explain the current snow accumulation and streamflow regimes, and their sensitivity to climate change. Two basins were selected in eastern Canada that represent contrasted landscape and meteorological conditions common in the region, namely Montmorency with a rugged forested landscape and cold/humid (boreal) climate, and Acadie with a flat agricultural landscape and warmer/less humid (hemiboreal) climate. These two contrasted basins were unique in the region, having both natural flow regimes and being sufficiently well instrumented to apply and validate a physically-based hydrological model (CRHM, Pomeroy et al., 2007). The two main questions addressed in this study are: (1) what are the differences and similarities between the current and future hydrological processes in these two contrasted basins? (2) What are the respective impacts of landscape and present-day climate conditions on present and future snow accumulation and streamflow regimes? This study is an attempt to advance efforts on understanding the relationship between current climate, landscape characteristics and basin response to climate change in humid cold regions.

2. Materials and methods

2.1. Study area and data

The Montmorency and Acadie basins are located in the St-Lawrence Valley in eastern Canada (Fig. 1). The Montmorency River Basin, which is the primary study area for this research, is mostly dominated by forests (85%) which are sub classified into mature coniferous forest (53%), mixed forest (6%) and regenerated forest (26%) (Fig. 1b). Forest clearings resulting from clear cutting and regeneration practices occupy about 8% of the basin area (Fig. 1b). Intensive farming activities take place in the Acadie River Basin, of which 77% is occupied by agricultural fields (Fig. 1c). Current and future hydrology of the Acadie were studied in a previous modelling study (Aygün et al., 2020b).

The Montmorency River Basin (Fig. 1b) differs from the Acadie River Basin (Fig. 1c) in terms of landscape and meteorological conditions (Table 1). Montmorency is dominated by forests on porous sandy loam over hilly uplands with varying altitudes, whereas the Acadie is occupied mostly by agricultural fields on compacted clay soils over flat lowlands (Table 1). Montmorency has a boreal climate (Dfc) (Peel et al., 2007) which is colder and more humid than the hemiboreal climate Dfb (Peel et al., 2007) of Acadie. The annual air temperature is about 6 °C higher in Acadie than in Montmorency (Table 1). Winters are considerably colder in Montmorency (−12.4 °C) than in Acadie (−6.6 °C). In spring, the air temperature is slightly below freezing point in Montmorency (−0.1 °C), while it is well above zero degree for Acadie (6.3 °C) (Table 1). Compared to the Acadie River Basin, the Montmorency River Basin receives more precipitation, and the snowfall ratio is almost twice that in Acadie (Table 1). Montmorency receives slightly more precipitation in winter than in spring, whereas it is the other way around for Acadie (Table 1).

The Montmorency River Basin encloses two densely studied watersheds, namely BEREV ("Bassin Bassin Expérimental du Ruisseau des Eaux-Volées") (20 km^2) managed by Laval University and Lac Laflamme (0.7 km^2) managed by Québec Ministry of Forests, Wildlife and Parks

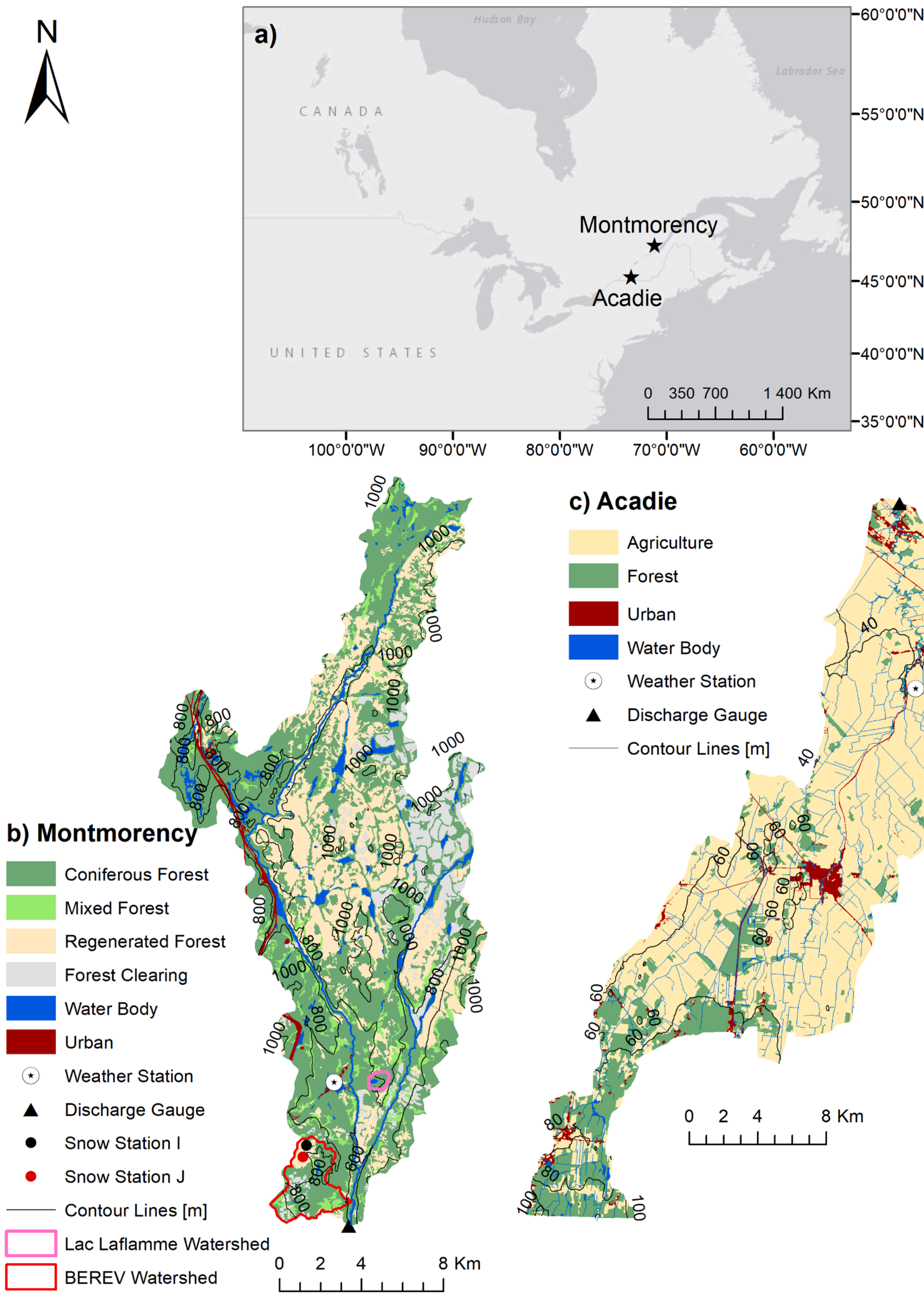

Fig. 1. Study area. a) Locations of Montmorency and Acadie River basins, b) Montmorency River Basin land cover, contour lines, discharge gauge, and main meteorological station, c) Acadie River Basin land cover, contour lines, discharge gauge, and main meteorological station. The Montmorency River Basin encloses the BEREV watershed with snow stations and the Lac Laflamme watershed with soil moisture/temperature stations.

(MFFP) (Fig. 1b). While earlier studies in the Lac Laflamme carried out modelling of snow accumulation and melt (Barry et al., 1990; Plamondon et al., 1984; Prévost et al., 1991), more recent studies focused on analyzing changes in soil water content and temperature (D'Orangeville et al., 2016; Houle et al., 2012) and also nutrient cycling (Houle et al., 2016). In the BEREV, earlier studies explored the impacts of forest harvesting on hydrological behaviour (Guillemette et al., 2005; Lavigne, 2007; Tremblay et al., 2008) and water quality (Tremblay et al., 2009), while more recent studies performed plot scale studies to examine catch efficiency of snowfall gauges (Pierre et al., 2019), energy and water budget (Isabelle et al., 2020), evapotranspiration (Hadiwijaya et al., 2020) and groundwater flow (Schilling et al., 2021).

Measurements of snow depth and density at snow stations I and J within the BEREV watershed (Fig. 1b) have been collected by the Québec Ministry of Environment and Fight against Climate Change (MELCC) and researchers from Laval University for the 2005–2019 period. Station I and Station J are located within regenerated forests with south-facing and north-facing slopes, respectively. Continuous measurements (since 1996) of soil temperature and moisture at the coniferous and mixed forest sites of the Lac Laflamme watershed

Table 1
Meteorological and landscape conditions of the Montmorency and Acadie River basins.

Characteristics	Montmorency	Acadie
Latitude	47° 25′ N	45° 11′ N
Longitude	71° 08′ W	73° 26′ W
Drainage area (km^2)	267	360
Elevation range (m)	550–1150	40–110
Slope range (°)	0–60	0–2
Dominant land cover	Forest (85%)	Agriculture (77%)
Dominant soil type (0–60 cm)	Sandy loam	Clay (compacted)
Saturated hydraulic conductivity (m/s)	6.2×10^{-6}	4.6×10^{-7}
Porosity	0.60	0.39
Climate type	Dfc (boreal)	Dfb (hemiboreal)
Mean annual temperature (°C)	1.3	7.2
Winter (DJF) temperature (°C)	−12.4	−6.6
Spring (MAM) temperature (°C)	−0.1	6.3
Annual precipitation (mm)	1460	1030
Winter precipitation (mm)	323	214
Spring precipitation (mm)	317	258
Annual snowfall ratio (%)	44	23
Mean annual relative humidity (%)	81	74
Mean annual wind speed (m/s)	1.9	3.0

(Fig. 1b) were acquired from the MFFP. Daily river discharge measured at the Forêt Montmorency gauge (ID: 051005) (Fig. 1b) were extracted from the database of Québec Center of Water Expertise for the 2005–2019 period.

Hourly temperature, wind speed, relative humidity and daily precipitation data have been acquired for the 2005–2019 period from the Forêt Montmorency weather station of Environment and Climate Change Canada (Fig. 1b). Daily precipitation observations were disaggregated uniformly over the hours of a day. The gaps in hourly and daily data, about 2% of the whole period, were filled with data from a MFFP weather station located in the Lac Laflamme watershed (Fig. 1b), 1.7 km away from the Forêt Montmorency weather station. The temperature was spatially distributed over the basin based on an environmental lapse rate of 5 °C/1000 m (Bergeron, 2016). Daily snowfall records were corrected for wind under-catch using the algorithm developed for the Alter-shielded Geonor gauge (Smith, 2007). The incoming shortwave radiation has been estimated using the method presented by Annandale et al. (2002) within the CRHM platform and validated against the reliable portion of existing data.

2.2. Modelling approach

Following Aygün et al. (2020b), the Cold Regions Hydrological Modelling (CRHM) platform (Pomeroy et al., 2007) was used to build a hydrological model for the Montmorency River Basin. Taking advantage of the object-oriented and modular structure of CRHM (Fang et al., 2013), relevant physical process modules were selected to simulate the dominant hydrological processes in the Montmorency River Basin (Table S1). The model was discretized into 78 hydrological response units (HRUs) that each represent one set of parameters and one control volume for mass and energy budgeting (Pomeroy et al., 2007; Zhou et al., 2014), and are defined by land cover, aspect, slope and elevation classes (see Supplementary Fig. S1). Building on earlier modelling results in Acadie, the model outputs from Montmorency were compared with previous results from the Acadie River Basin (Aygün et al., 2020b), updated for this study.

Subsurface flow and groundwater-surface interactions in Montmorency River Basin were simulated using physically based parameters and principles on hillslopes, using the Hillslope module in CRHM (Table S1). This module is a modified version of the soil moisture balance modules in CRHM (Dornes et al., 2008; Fang et al., 2010), comprising a near-surface detention layer over two soil layers and a groundwater layer with provision for macropore flow between sub-surface layers and from surface to groundwater (Fang et al., 2013). The near-surface detention layer allows the surface runoff to flow through a porous medium as a transient flow pathway, which allows handling the water storage in loose organic material in the forest floor (Pomeroy et al., 2016). Horizontal and vertical flows from soil layers and groundwater layer are calculated based on Darcy's law, where Brooks and Corey's relationship (Brooks and Corey, 1964) is used to estimate the actual hydraulic conductivity in the unsaturated zone.

Estimation of the parameters in the Montmorency River Basin was performed based on the well-studied BEREV and Lac Laflamme watersheds (Section 2.1; Fig. 1b) and also other snow-dominated basins with similar landscape characteristics. The soil profile parameters were estimated from previous studies performed at Lac Laflamme (Barry et al., 1988; Jutras, 2012; Ouimet and Duchesne, 2005). Vegetation height and stalk diameter were obtained from the ecoforest maps produced in southern Québec by the MFFP. Mature coniferous and mixed forest stands were assigned to have a stalk diameter of 60 cm and canopy height of 14 m, whereas regenerated forest HRUs were assigned a stalk diameter of 40 cm and a canopy height of 6 m. The Leaf Area Index (LAI) was set to be 2.9 m^2/m^2 and 3.4 m^2/m^2 for regenerated and mature balsam fir forests, respectively, as reported by previous studies for the BEREV subbasin of Montmorency (Hadiwijaya et al., 2020; Isabelle et al., 2020; Parajuli et al., 2020). The maximum canopy snow load capacity was set to 3.3 kg/m^2 and 6.3 kg/m^2 for regenerated forest and mature coniferous forest, respectively. These values are transferred from the studies performed in boreal forests of western Canada (Hedstrom and Pomeroy, 1998; Pomeroy et al., 2012). All the HRUs were routed to the streamflow network, where routing lengths were calculated as median distances from each HRU to the closest tributary. The model was run at an hourly time step. Evaluation of the CRHM model performance for SWE, soil moisture and river discharge in Montmorency River Basin is provided in the Supplementary material (Figs. S2–5). Hourly simulations for Acadie are available for 1996–2019, and detailed model setup and results are given in Aygün et al. (2020b).

2.3. Perturbed climate

The potential impacts of climate change on the hydrology of the Montmorency River Basin were explored via climate sensitivity analyses, where the historical long-term temperature and precipitation data (2005–2019) were perturbed according to the ensemble of climate projections produced by Charron (2016). These projections were generated from a set of 11 downscaled global climate simulations provided by the Coupled Model Intercomparison Project (CMIP5) (Charron, 2016). They are produced for two greenhouse gas emission scenarios (RCP 4.5 and RCP 8.5) and for two periods (2041–2070 and 2071–2100) for all the administrative regions in the province of Québec (Charron, 2016). Based on these scenarios, temperature warming up to 8 °C (0–8 °C at 1 °C degree interval) and an increase in total precipitation ("wetting") up to 20% (0–20%, 5% interval) were considered in the sensitivity analyses. These changes thus encompass the spread in projections for the mid- and end of century under the RCP 4.5 and RCP 8.5 scenarios for the administrative region where the Montmorency River Basin is located (Fig. 1).

The climate sensitivity analyses carried out previously for the Acadie River Basin (Aygün et al., 2020b) were repeated by updating the reference climate period from 1996–2019 to 2005–2019 in order to match the reference climate period used in Montmorency, which had a shorter observational record. Therefore, the results presented for the Acadie River Basin in this paper are modified from those presented in Aygün et al. (2020b).

The hydrological sensitivity of the Montmorency River Basin (Fig. 1a) was compared with that of the Acadie River Basin (Fig. 1b) by assessing the changes in snow and water fluxes using sensitivity response surfaces. Detailed comparisons were carried out for the 3 °C and 6 °C warming scenarios, which respectively represent the mean warming projection for the 2041–2070 and 2071–2100 periods under the high emission scenario for both basins (Charron, 2016). These warming scenarios were modulated with a minimum (0%) and maximum (20%) increase in precipitation based on the range of existing scenarios in order to analyze the potential compensation effect of increasing precipitation on snow and water fluxes impacted by warming. The reference run ($\Delta t = 0\,^{\circ}C$ & $P = 100\%$) represents the historically averaged observed data for both basins over the 2005–2019 period.

2.4. Permuted baseline climate experiments

In order to explore the respective roles of present-day climates and basin landscape on the climate sensitivity of hydrological responses, a set of climate sensitivity analyses were performed in which the historical climates of both basins were permuted. Historical (2005–2019) time series of air temperature and precipitation of the Acadie River Basin were thus used as meteorological inputs for the Montmorency River Basin and vice-versa. This resulted in climate sensitivity analyses for four distinct combinations of present-day climate and landscapes: Acadie-type landscape under its own (hemiboreal) climate, Acadie-type landscape under Montmorency (boreal) climate, Montmorency-type landscape under its own (boreal) climate and Montmorency-type landscape under Acadie (hemiboreal) climate. The climate sensitivities of hydrological variables, i.e., relative changes (%) in hydrological variables in response to changes in temperature and precipitation, were compared for the most likely climate change scenarios for the 2041–2070 period: 3 °C warming and/or 20% increasing precipitation.

3. Results

The model performance in simulating daily streamflow at the basin outlet for the 2005–2019 is slightly better in Montmorency with a Kling-Gupta Efficiency (KGE, Gupta et al., 2009) of 0.82 and a mean bias of −0.6% (Fig. S5a and c) than in Acadie with a KGE of 0.69 and a mean bias of 1.8% (Aygün et al., 2020b).

3.1. Simulated historical water fluxes

Montmorency receives 40% higher annual precipitation (1460 mm) than Acadie (1030 mm) (Fig. 2a and b). Almost half (44%) of the annual precipitation in Montmorency occurs as snowfall, whereas the snowfall ratio in Acadie is 23% (Fig. 2a and b). In the southernly and warmer Acadie, evapotranspiration and streamflow are the two main outfluxes of the annual water balance, each accounting for 45% of annual precipitation. In Montmorency, mean annual streamflow (1067 mm; 73% of annual precipitation) is the largest outflux, with only 18% of the annual precipitation being lost through evapotranspiration (Fig. 2a and b). Groundwater is an important part of the hydrological system in Montmorency, which sources 48% (513 mm) of the streamflow as baseflow. Annual sublimation losses in Montmorency (140 mm) reach ~22% of the annual snowfall, compared with ~14% (29 mm) in Acadie.

In both basins, snowmelt generates a substantial proportion (Montmorency: 46%, Acadie: 42%) of the mean annual streamflow (Fig. 2c and d). The snowmelt contribution to annual streamflow is marked by the sharp increase in cumulative runoff in both basins during snowmelt periods (May–June in Montmorency and April–May in Acadie). The mean simulated annual peak SWE in Montmorency (405 mm) (Fig. 2c) is about six times greater than that in Acadie (65 mm) (Fig. 2d). Under historical climate conditions, the annual peak SWE in Montmorency occurs on April 10 (Fig. 2c), which is more than a month later that in Acadie where it presently occurs in early March (Fig. 2d).

3.2. Responses of snow fluxes to warming

The ratio of snowfall to total precipitation in both basins, shown by percentages on panel a and b of Fig. 3, exhibits a decrease in response to warming. Under 6 °C warming, only 10% of the total precipitation occurs as snowfall in Acadie (Fig. 3b), whilst the snowfall ratio in Montmorency drops to 23% (Fig. 3a), which is equal to the snowfall ratio in Acadie under reference climate conditions (Fig. 3b). The amount of snowmelt in both basins is lower under warmer temperatures, due to these decreased snowfall ratios (Fig. 3a and b).

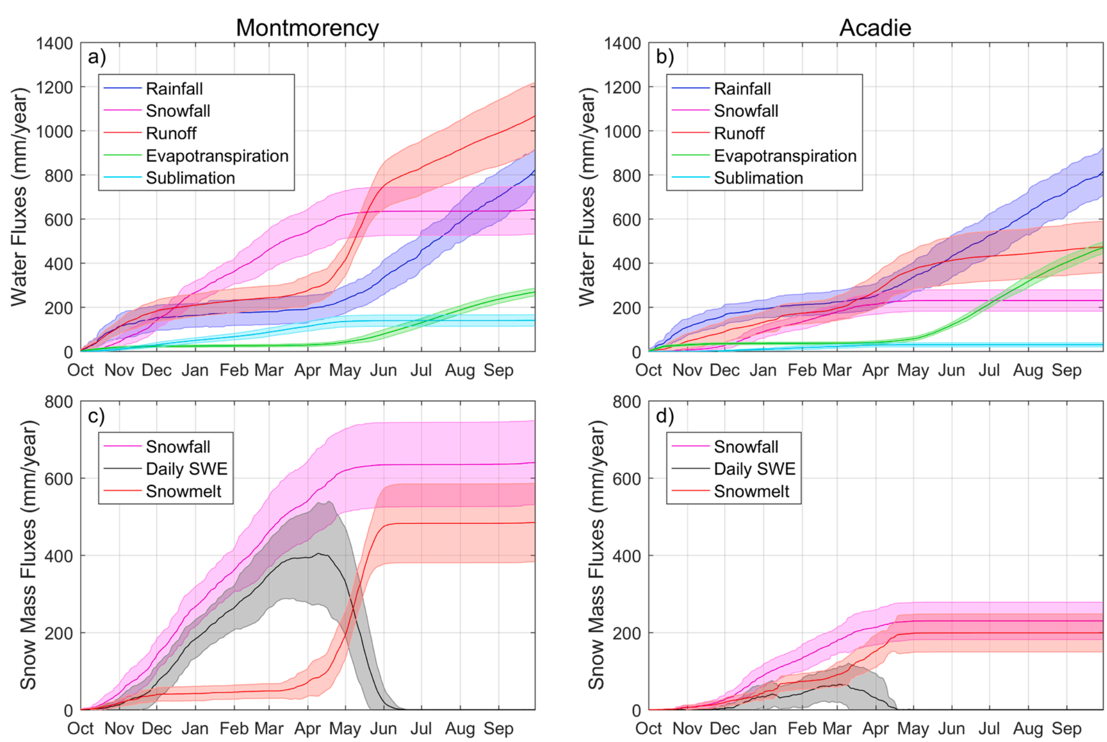

Fig. 2. Average annual cumulative water fluxes in a) Montmorency and b) Acadie, and average cumulative snow fluxes and daily SWE in c) Montmorency and d) Acadie over the 2005–2019 period. The shades around the average values represent the inter-annual variability (±standard deviation).

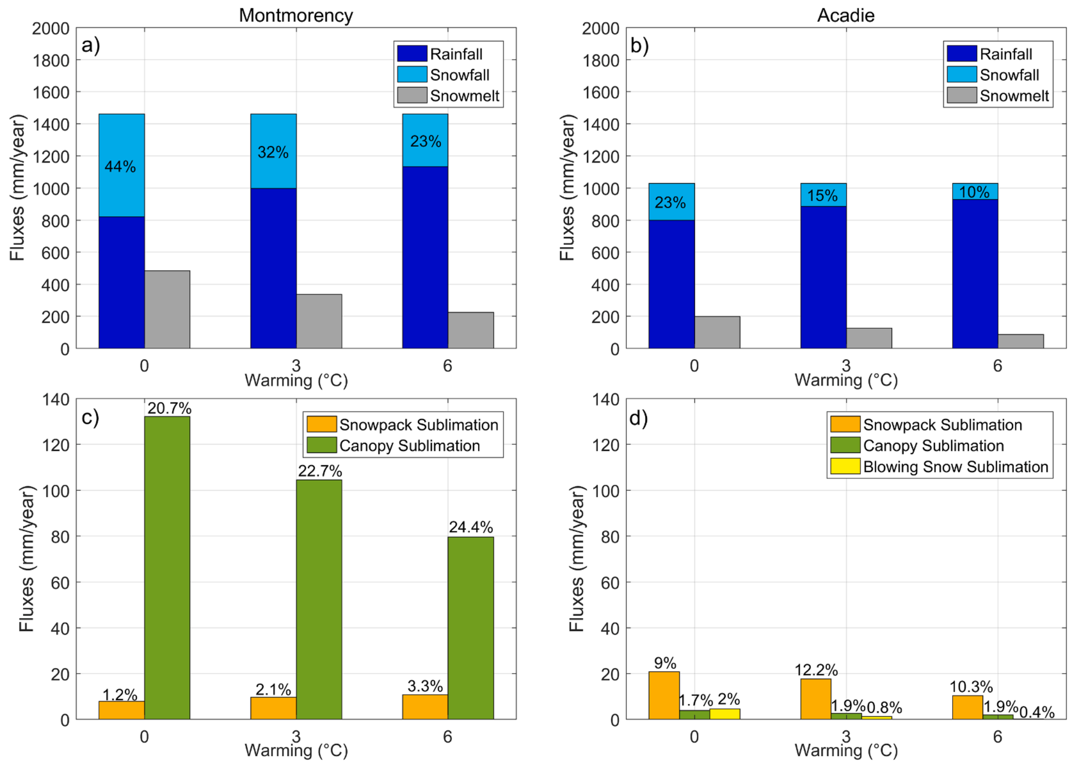

Fig. 3. Rainfall, snowfall, snowmelt, and sublimation losses under selected climate change scenarios in a–c) Montmorency River Basin, and b–d) Acadie River Basin. The ratio of annual snowfall to total precipitation is given on each bar on panel a and b. The ratios above each bar on panel c and d represent the ratio of sublimation to annual snowfall.

Intercepted snow sublimation removes a considerable amount of snow (132 mm/year; 21% of the annual snowfall) in Montmorency under reference climate conditions (Fig. 3c). This is because Montmorency is dominated by coniferous trees (Fig. 1b and Table 1) that intercept a sizeable fraction of seasonal snowfall, which subsequently sublimates over the long winter. Warming air temperatures in Montmorency reduce the total sublimation from intercepted snowfall to 104 mm and 79 mm with 3 °C and 6 °C warming, respectively (Fig. 3c), which can be explained by the reduced snowfall ratio and increased unloading and drip of intercepted snow from the forest canopy (Fig. 3a).

In comparison with the canopy sublimation, sublimation from the snowpack surface is very low, reaching only 1.2% of the annual snowfall in Montmorency under reference climate conditions (Fig. 3c). In Acadie, on the other hand, annual sublimation is mostly composed of snowpack sublimation (20.8 mm/year; 9% of the annual snowfall), whereas the canopy sublimation loss (4 mm/year; 1.7% of the annual snowfall) is the smallest term of the snow mass balance under the reference climate conditions (Fig. 3d). These contrasted sublimation losses can be explained with the landscape of Acadie, which is dominated by well-exposed open agricultural fields (Fig. 1c and Table 1) that are subjected to high winds and relatively high surface sublimation rates, while only 17% of the basin is covered by forest (of which 60% is deciduous), therefore resulting in relatively small canopy sublimation losses at the basin scale.

Compared to the reference climate conditions, the snowpack sublimation to snowfall ratio is higher in both basins under warmer temperatures (Fig. 3c and d), which can be explained by the greater available energy for sublimation under warmer temperatures. Canopy sublimation ratios are also greater, particularly in Montmorency under warmer temperatures (Fig. 3c). These higher sublimation ratios suggest that sublimation is a more efficient snow removal process under warmer temperatures. Since relative humidity and wind speed remain unchanged in the climate sensitivity experiments, higher sublimation to snowfall ratio implies that the saturated vapor pressure at the snow/air interface increases more on average than the atmospheric vapor pressure, leading to a greater water vapour deficit and hence higher sublimation fluxes.

While there is no blowing snow sublimation component in the forested Montmorency as blowing snow transport is suppressed by the tall vegetation, blowing snow sublimation reaches 2% of the annual snowfall under reference climate conditions in more open Acadie (Fig. 3d). In comparison with the other sublimation components in Acadie, blowing snow sublimation shows the greatest sensitivity to warming, declining by 74% and 90% with 3 °C and 6 °C warming, respectively (Fig. 3d). This is due to the increasing inter-crystal bond strength and cohesion of snow as it warms, which raises the threshold wind speed required to initiate saltation (Li and Pomeroy, 1997).

3.3. *Altered snow regimes in response to climate change*

The annual peak SWE (the mean annual maximum SWE) decreases in response to warming in both basins (Fig. 4a and d), with more dramatical changes simulated for Acadie. While the peak SWE declines by about 10% per °C warming in Montmorency (Fig. 4a), the peak SWE in Acadie decreases by 60% per °C warming (Fig. 4d). The lower SWE sensitivity in the Montmorency is due to the colder temperatures in this region compared to the Acadie (Table 1).

Under 3 °C warming, which represents the mean warming projection for the mid-century for both basins, the decline in peak SWE in Acadie (79%) is twice as great as in Montmorency (38%) (Fig. 4a and d). Under this scenario, a 20% increase in precipitation would compensate 48% of the decline in peak SWE in Montmorency, but only 13% in Acadie (Fig. 4a and d). A 6 °C warming, which represents the mean warming for the 2071–2100 period under the high emission scenario, reduces peak SWE in Montmorency by 70%, from 405 mm to 115 mm (Fig. 4a and d), which is still greater than the peak SWE in Acadie under reference climate conditions (65 mm) (Fig. 2d).

The annual peak SWE shifts towards earlier dates under almost every

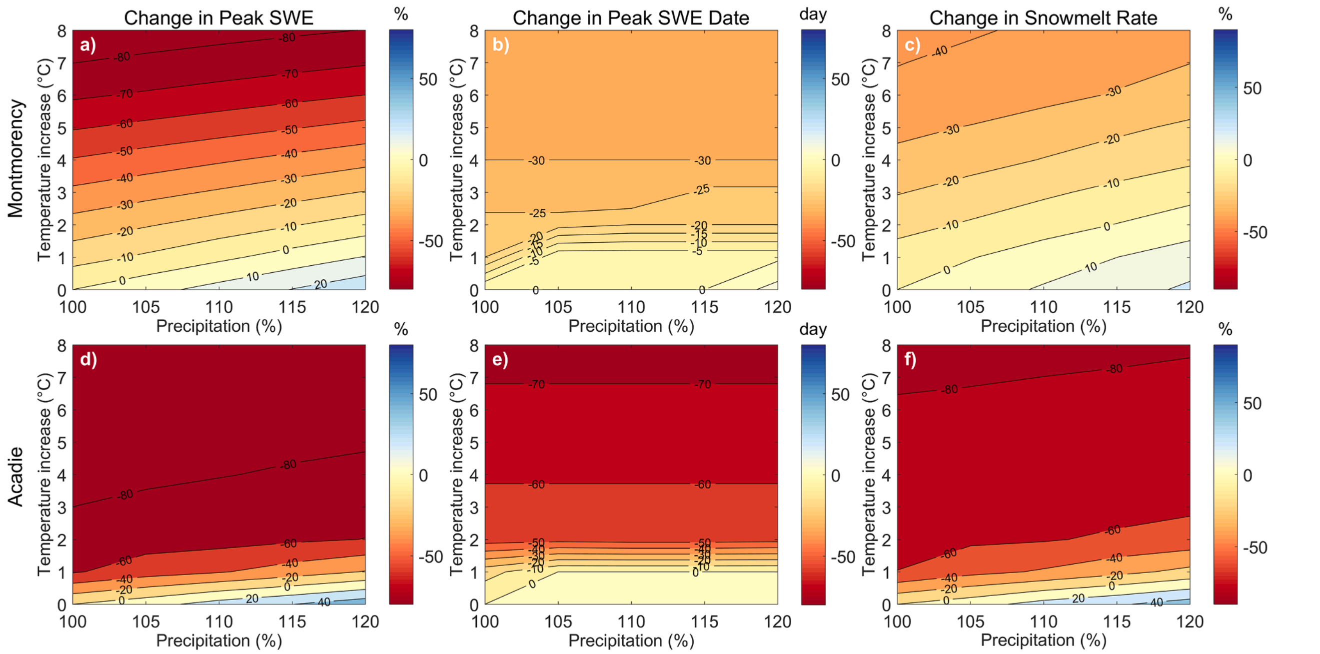

Fig. 4. Sensitivity of snow accumulation to changing climate in Montmorency and Acadie. Changes in annual peak SWE in Montmorency (a) and Acadie (d); changes in annual peak SWE date in Montmorency (b) and Acadie (e); relative changes in snowmelt rate in Montmorency (c) and Acadie (f).

warming scenario in both basins, with greater shifts simulated for Acadie. For instance, peak SWE advances by less than a month in Montmorency under 3 °C warming, while the same scenario causes the peak SWE in Acadie to shift by almost two months (Fig. 4b and e). The peak SWE timing changes the fastest within the 1–2 °C warming band in both basins, but with comparatively fast changes in peak SWE only in Acadie.

Snowmelt rates are slower under all warming scenarios in both basins as long as there is no increase in precipitation (Fig. 4c and f). Like the peak SWE response surface patterns (Fig. 4a and d), snowmelt rates decline uniformly by 10–15% per °C warming in Montmorency, whereas in Acadie snowmelt rates decrease faster, reaching 60% for a 1 °C warming. This is because warming leads to shallower snowpacks which melt out earlier in the year when the available energy is lower, therefore

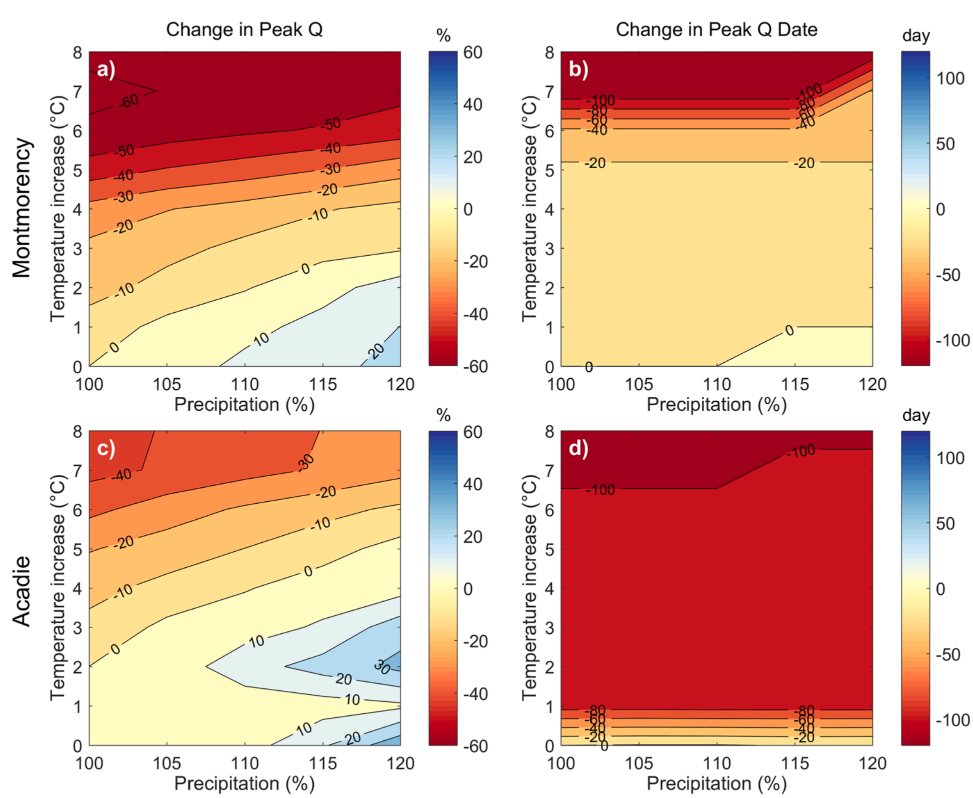

Fig. 5. Sensitivity of streamflow to changing climate in the Montmorency and Acadie River basins. Changes in annual peak streamflow in Montmorency (a) and Acadie (c); changes in annual peak streamflow timing in Montmorency (b) and Acadie (d).

leading to overall slower melt rates. Fig. 4c and f also suggest a possibility for faster snowmelt rates in the future when limited warming (<1°C for Acadie, and <2.5 °C for Montmorency) is accompanied by increasing precipitation (Fig. 4c and f). The faster melt rates could reflect the higher incoming energy available for deeper snowpack under wetting and limited warming scenarios.

3.4. Comparison of peak streamflow sensitivities

The annual peak discharge (the mean annual peak discharge) could increase by up to 25% and 40% in Montmorency and Acadie, respectively, depending on the warming and wetting scenario (Fig. 5a and c). This zone of positive peak discharge sensitivity is delineated by the 0% contour in Fig. 5a and c, below which the peak streamflow shows an increase. In Montmorency, a 20% increase in precipitation could compensate peak flow declines due to warming up to 2.9 °C. However, beyond +3 °C, warming impacts predominate and peak flow decreases, which can be attributed to the declining peak snow accumulation (Fig. 4a). Fig. 5c indicates a two-layer positive sensitivity zone for Acadie. While the increasing peak flow in the first layer (warming <1 °C) in Fig. 5c is attributed to the higher snow accumulation in response to increasing precipitation and limited warming (<1 °C, see Fig. 4d), the second layer (warming >1 °C) can be explained by increasing peak flows under a mixed rain and snowmelt winter regime, as explained next.

Warming causes a shift in peak streamflow timing towards earlier dates in both basins, with greater shifts simulated for Acadie (Fig. 5b and d), which is in line with the earlier peak SWE occurrence under warming scenarios (see Fig. 4b and e). The peak streamflow in Acadie shifts by 80 days earlier in response to 1 °C warming (Fig. 5d) and occurs before the peak SWE date under the same warming level (Fig. 4e). This finding suggests that the peak flow in Acadie desynchronizes from the peak snow accumulation in response to a 1 °C warming, i.e. shifting towards a rain-dominated flood regime. For Montmorency, the timing of peak streamflow is relatively less sensitive to warming compared to the timing of peak SWE, particularly for a warming between 0 °C and 5 °C (Figs. 5b vs 4b). However, beyond 6 °C warming, the peak streamflow timing in Montmorency becomes very responsive, advancing by roughly-four months when warming reaches 7 °C. Under these conditions, the streamflow regime of Montmorency shifts towards a mixed snowmelt/rainfall regime with peak flows occurring in winter.

Under the mid and late 21st century climate conditions where warming exceeds 3 °C, the Acadie River mean hydrograph becomes very flashy and the seasonality of precipitation dictates the magnitude and timing of the annual peak streamflow (Fig. 6c); in other words, the flow regime of the Acadie River Basin transits to a rainfall dominated regime. In contrast, the Montmorency River conserves a distinct snowmelt-dominated peak streamflow under 3 °C warming. Moreover, a 20% increase in precipitation almost completely (94%) counterbalances the decline in peak streamflow caused by a 3 °C warming; however, the peak flow occurs 19 days earlier, i.e., on April 23 rather than May 9 (Fig. 6a). This highlights the considerable uncertainty in future peak streamflow magnitude and timing and flood risks caused by the uncertainties in projected precipitation. Meanwhile, a 6 °C warming causes the flow regime of Montmorency River to transit from a snowmelt to a mixed snowmelt/rainfall regime (Fig. 6a).

Warmer temperatures cause an increase in winter flows in both basins (Fig. 6a, c), which is due to the increase in available water in winter due to higher winter rainfall and more frequent mid-winter snowmelt events. Increasing precipitation leads to even higher streamflow in winter. This is also evident in the increase in flows with exceedance probabilities between 0.3 and 0.8, and 0.5 and 1, respectively for Acadie and Montmorency (Fig. 6b and d). In Acadie, winter becomes the active flood season (Fig. 6c), due to abundant rainfall which combines with winter snowmelt and restricted infiltration over frozen ground.

3.5. Relative influence of climate and basin landscape on the hydrological sensitivity to climate change

The respective roles of the landscape features and current climates on the climate sensitivity of key hydrological variables are explored in Figs. 7–10. In these figures, the vertical double arrows indicate the effect of regional climate, while the horizontal double arrows indicate the influence of the basin landscape on the climate sensitivity of a given hydrological variable. The reference (historically averaged) baseline values of hydrological variables under a given regional climate and landscape combination are given in parentheses in panel a of Figs. 7–10.

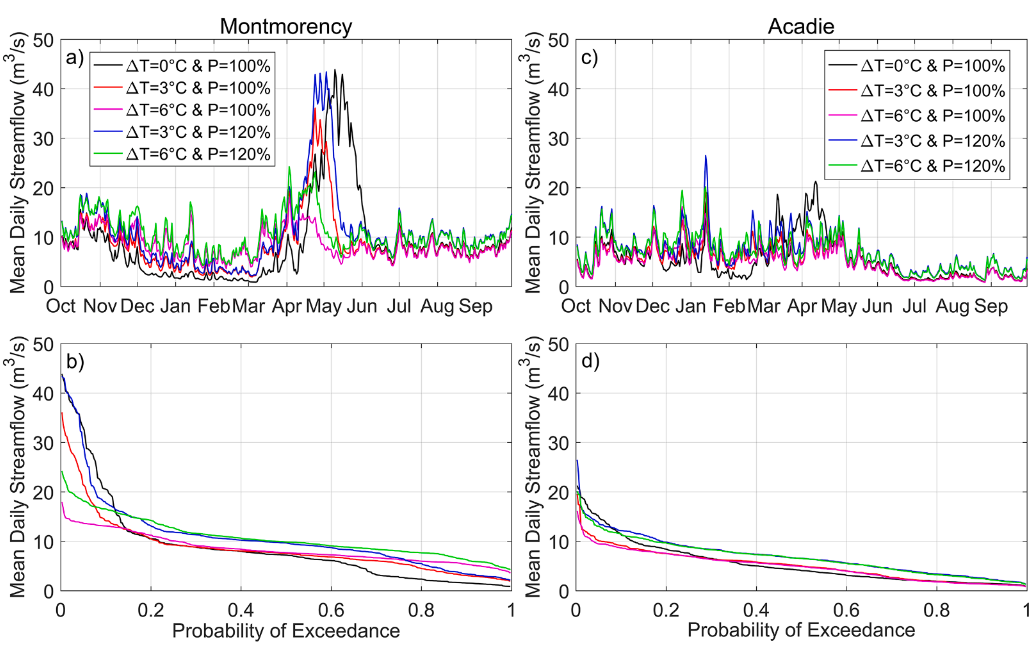

Fig. 6. Changes in mean daily streamflow and exceedance probability of mean daily streamflow to selected warming and increasing precipitation scenarios in (a, b) Montmorency River Basin and (c, d) Acadie River Basin.

3.5.1. Peak SWE

Focusing on the impact of regional climate on the current peak SWE (vertical arrows in Fig. 7a), the snowier and colder boreal climate of Montmorency leads to greater peak snow accumulation compared to the hemiboreal Acadie climate, under both Montmorency-type (405 mm > 67 mm) and Acadie-type (431 mm > 65 mm) landscapes. On the other hand, the sensitivity of peak SWE to warming and wetting is found to be much more pronounced when both landscapes are forced by the warmer and drier hemiboreal baseline climate, than by the colder and more humid boreal climate (vertical arrows in Fig. 7a–c). This highlights the strong dependence of the climate sensitivity of the peak SWE to current climate conditions. Under hemiboreal climate conditions, a small to moderate warming leads to significant declines in snow accumulation due to already milder temperatures (Table 1).

Looking at the influence of landscape on the current peak SWE (horizontal arrows in Fig. 7a), the simulated peak SWE in Montmorency-type landscape (405 mm) is smaller than in Acadie-type landscape (431 mm) when both are forced by the boreal climate. This can be explained by the higher (canopy) sublimation losses in the forested landscape of Montmorency (Fig. 3c). When forced by the warmer and rainier hemiboreal climate, canopy sublimation losses in Montmorency decrease but the total sublimation ratio in Montmorency (36%) remains greater than that in Acadie (13%), because the greater canopy storage in coniferous forests favour sublimation losses. However, the amount of snowmelt simulated in mid-winter in Montmorency is about 22% less than that in Acadie, mostly due to the reduced amount of energy available for melting in Montmorency due to shading by the forest canopy. This reduced snowmelt compensates the sublimation losses and as a result, the peak snow accumulations are almost equal in both basins when forced by the hemiboreal climate (Fig. 7a). Likewise, the peak SWE shows rather similar sensitivities to warming and wetting in both basins when forced by the same climate (Fig. 7a–c). These results show that the climate sensitivity of the peak SWE is little influenced by the landscape but is rather shaped by the current regional climate condition.

3.5.2. Water fluxes

Present-day mean annual effective precipitation was calculated as the sum of snowmelt and net rainfall, where net rainfall is the gross rainfall minus interception losses. Hence, differences in baseline mean effective precipitation and their climate sensitivities reflect the varying influences of interception losses (canopy sublimation and evaporation) and snowpack sublimation losses. Present-day effective precipitation (values in parentheses in Fig. 8a), are higher under Montmorency (boreal) baseline climate due to its higher precipitation rates and cooler and more humid climate (Fig. 3a vs b).

The Montmorency-type landscape enhances the climate sensitivity of the effective precipitation, especially under hemiboreal baseline climate forcing (Fig. 8a–c). In order to quantify the influence of vegetation on effective precipitation sensitivity, the Montmorency basin was artificially deforested in the CRHM model, keeping the other basin properties, including soils, unchanged. The results showed that sensitivity of effective precipitation in the deforested Montmorency decreased to 20–21% in response to combined 3 °C warming and 20% wetting, which is lower than for the forested Montmorency-type landscape (23–27%) and almost the same as for the Acadie-type landscape (20–21%) (Fig. 8c). These findings highlight that canopy interception plays the key role in modifying the effective precipitation sensitivity to climate change.

Snow storage on the canopy in Montmorency is particularly prone to sublimation losses, so that the shift from snowfall to rainfall in a warmer climate decreases the canopy losses and leads to a slight increase in mean effective precipitation (Fig. 8a), even if canopy and snowpack sublimation themselves become more efficient under a warmer climate (i.e., sublimation ratios increase, see Fig. 3c). This suggests that sublimation from the canopy is a more efficient process to remove water from this basin than is evaporation of rainfall stored on the canopy, due to the longer residence time of snow on the canopy compared to rainfall.

The fraction of the mean effective precipitation that infiltrates in the soils varies among the two landscapes and climates. The "buffering capacity" is defined similarly to Herron and Wilson (2001) and Van Tiel et al. (2021), as the ratio of the annual mean infiltration to the mean effective precipitation, which describes the capacity of a catchment to absorb surface water fluxes and its ability to attenuate runoff and enhance baseflow. This buffering capacity is lower in the Acadie-type landscape (0.40–0.59) than in the Montmorency-type landscape (0.54–0.73) when both are forced by the same climate (Fig. 9a). This suggests that the Acadie-type landscape is more efficient in translating available water to runoff than Montmorency-type landscape, due to its compacted agricultural soils that have a lower infiltration and storage capacity. The more porous forested soils of Montmorency, on the other hand, favour infiltration (Fig. 9a, b).

Artificially deforesting Montmorency has a small impact on its buffering capacity (Fig. 9a), which confirms that soils predominantly affect the buffering capacity of Montmorency. The buffering capacity of both basins is also higher when forced by the warmer and less snowy hemiboreal climate than by the cooler and wetter boreal climate (Fig. 9a, b), which suggests that infiltration decreases overall under a snowier climate, because the soil remains saturated and/or frozen for a long period during spring snowmelt and restrict infiltration. The buffering capacity of Montmorency-type landscape increases in response to warming and wetting, especially under boreal climate forcing (Fig. 9b, c), reflecting the larger snowfall to rainfall conversion for this climate (Fig. 3a, b) and decreased influence of snowmelt on soil saturation.

The correlation coefficient between average monthly discharge and precipitation (Corr(Q/P): Fig. 10a), provides information on the current seasonal synchronicity between discharge and precipitation for each

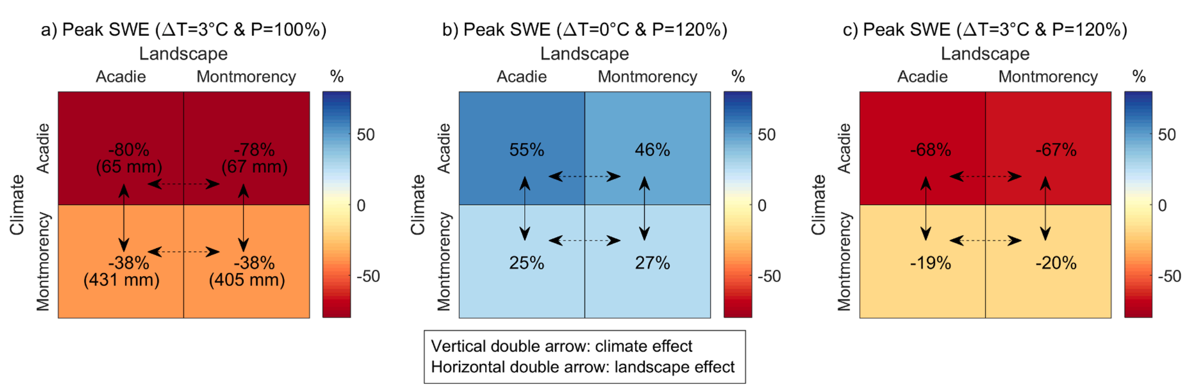

Fig. 7. Influence of basin landscape and climate conditions on the climate sensitivity of the annual peak snow water equivalent (SWE) from permuted baseline climate experiments. (a) 3 °C warming; (b) 20% wetting; (c) 3 °C warming with 20% wetting. The climate sensitivity (%) is displayed by the color scale and baseline conditions (mm) are shown in parenthesis in panel a.

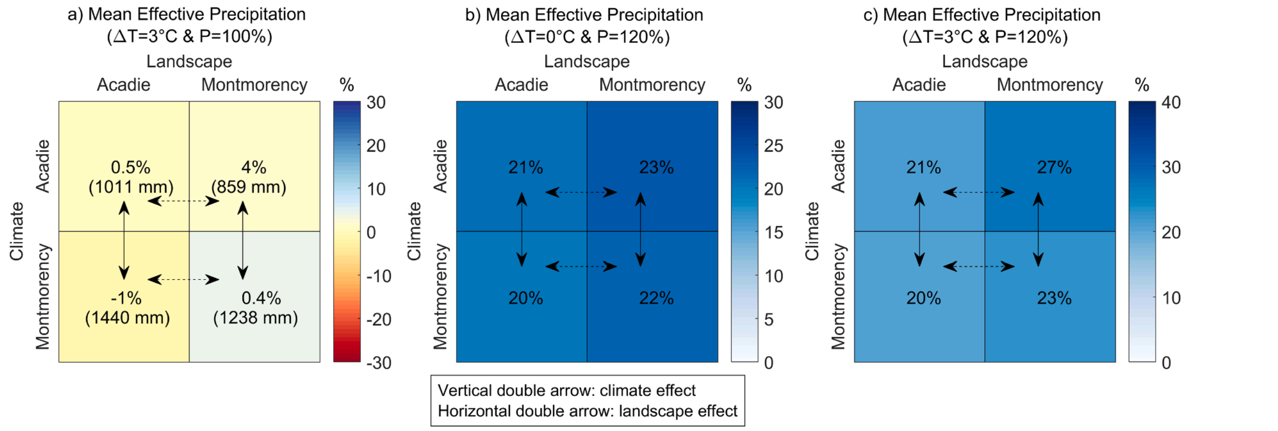

Fig. 8. Influence of basin landscape and climate conditions on the climate sensitivity of the effective precipitation from permuted baseline climate experiments. (a) 3 °C warming; (b) 20% wetting; (c) 3 °C warming with 20% wetting. The climate sensitivity (%) is displayed by the color scale and baseline conditions (mm) are shown in parenthesis in panel a. Sensitivity values for the artificially deforested Montmorency basin are presented in the text of Section 3.5.2.

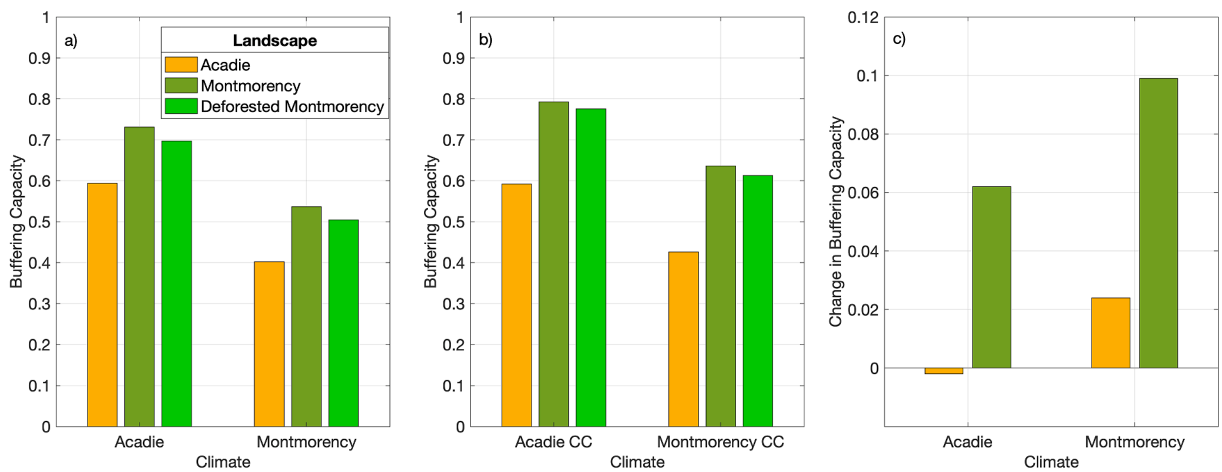

Fig. 9. Influence of basin landscape and climate conditions on the buffering capacities. (a) Buffering capacity of Acadie, Montmorency and deforested Montmorency under current climate conditions; (b) Buffering capacity of Acadie, Montmorency and deforested Montmorency under combined 3 °C warming and 20% increased precipitation climate change (CC) scenario; (b) Change in buffering capacity of Acadie and Montmorency in response to 3 °C warming and 20% increased precipitation.

combination of climate and landscape. The basins exhibit much lower Corr(Q/P) (0.07–0.14), i.e., a lower synchronicity between P and Q, when forced by the baseline boreal climate (Montmorency) than when forced by the hemiboreal (Acadie) climate (Corr(Q/P): 0.28–0.66) (Fig. 10a). This is because the colder boreal climate has a higher snowfall fraction (44%) than the hemiboreal climate (23%) (Fig. 3a and b), which results in larger snow accumulation (storage) that delays discharge until late spring. This snow damming effect is very climate sensitive in many basins around the world and its loss is an indicator of decoupling between snowpack and streamflow regimes (López-Moreno

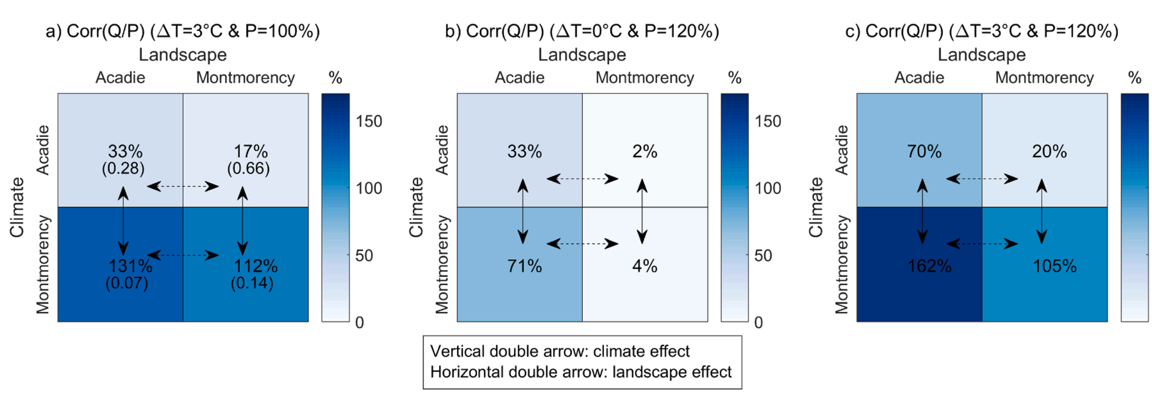

Fig. 10. Influence of basin landscape and climate conditions on the climate sensitivity of the precipitation-discharge synchronicity (Corr(Q/P)) from permuted baseline climate experiments. (a) 3 °C warming; (b) 20% wetting; (c) 3 °C warming with 20% wetting. The Corr(Q/P) under baseline climate conditions is displayed in parenthesis in panel a.

et al., 2020).

The coupling between P and Q increases under all climate change scenarios and climate/landscape combinations (Fig. 10a–c), with the largest increases simulated under a combined warming and wetting scenario (Fig. 10c). The boreal (Montmorency) climate enhances the climate sensitivity of Corr(Q/P), i.e., the relative increase in the seasonal synchronicity between precipitation and streamflow is more pronounced in this climate. However, although the climate sensitivity of Corr(Q/P) is higher under boreal climate forcing, the absolute Corr(Q/P) values themselves remain expectedly higher under the rainier hemiboreal climate forcing, suggesting a more synchronized Q and P.

The Montmorency-type landscape favors larger present-day Corr(Q/P) under the same baseline climate conditions, especially under the warmer and rainier Acadie climate (Fig. 10a). This can be explained by two factors: (i) the higher slopes in Montmorency which favor more rapid transfers of surface runoff; (ii) the higher infiltration and storage capacity that lead to a smoother hydrograph in Montmorency compared to Acadie (see Fig. 6); the lag-1 autocorrelation of monthly streamflow is indeed correspondingly higher for the Montmorency-type landscape (0.21) than for the Acadie-type landscape (0.09), which would act to increase Corr(Q/P) (Fig. 10a).

To isolate the effect of slopes from soil effects, CRHM model parameters describing HRU slopes were artificially lowered for Montmorency to approximately match those in Acadie. This resulted in a Corr(Q/P) of 0.22 when forced by the Acadie climate, which is lower than the original value (0.66 in Fig. 10a) and more comparable to the Acadie-type landscape under Acadie climate (0.28 in Fig. 10a). The higher slopes in Montmorency thus appear to explain the current-day tighter coupling between Q and P, especially under a rainier climate. Under a changing climate, however, the Montmorency-type landscape leads to smaller increases in the synchronicity between P and Q (Corr(Q/P)) compared to the Acadie-type landscape (Fig. 10a–c). This can be explained by the greater increases in the buffering capacity of the Montmorency-type landscape under climate change, as described in the previous section (Fig. 9c).

3.5.3. Peak discharge

The response of peak discharge to warming and wetting is illustrated in Fig. 11. Present-day peak discharges are higher under the boreal climate (0.16–0.24 $m^3/s/km^2$) than under the hemiboreal climate (0.05–0.06 $m^3/s/km^2$) (Fig. 11a) as the current peak SWE is greater and the buffering capacity is lower, for the boreal climate (Figs. 7a and 9a). Interestingly, streamflow regimes show less sensitivity to warming (Fig. 11a) than do snowpack regimes (Fig. 7a) under any baseline climate and landscape conditions.

When the slopes are artificially lowered in Montmorency-type landscape to match those in Acadie, the present-day peak discharge decreases by about 20% under both boreal climate (0.13 $m^3/s/km^2$) and hemiboreal climate (0.04 $m^3/s/km^2$). This means that higher slopes lead to higher peak discharges. Yet the Acadie-type landscape, which has lower slopes, exhibits higher present-day peak discharges than the Montmorency-type landscape under both climates (Fig. 11a), due to its lower buffering capacity (Fig. 9a).

Peak Q shows a stronger sensitivity to warming under the boreal climate (−15% to −18%) than under the hemiboreal climate conditions (−8% to −9%) (Fig. 11a). This is opposite to the much greater sensitivity of peak SWE seen under the hemiboreal climate compared to the boreal climate, in response to the same level of warming (Fig. 7a). This dichotomy occurs because when the hemiboreal climate is warmed by 3 °C, the peak Q decouples from the snow cycle (Fig. 5d), so that the large declines in peak SWE (Fig. 7a) do not translate into large changes in peak Q (Fig. 11a).

While warming alone leads to an overall decline in peak discharge, combined warming and wetting results in higher peak discharges for all climate and landscape combinations, except for the Montmorency-type landscape under its own climate (Fig. 11c). The hemiboreal climate shows a greater sensitivity (37%–47%) of peak flow to wetting than the boreal climate (18%–24%) (Fig. 11b), which can be explained by the greater increases in peak SWE under the hemiboreal climate in response to wetting (Fig. 7b). Similarly, the significant declines in peak SWE in response to combined warming and wetting of the hemiboreal baseline climate, (Fig. 7c) are not reflected by decreasing peak flows, which rather rise by 20–24% (Fig. 11c). This is due to increased precipitation in the form of rainfall leading to increased annual peak runoff (net rainfall + snowmelt – infiltration) (Supplementary Fig. S6f). In contrast, when forced with the boreal baseline climate, the peak Q remains synchronized with the peak snowmelt timing under 3 °C warming (Fig. 5b), and the peak Q of both basins responds more strongly to changes in peak SWE even though the declines in peak SWE are smaller (Fig. 7a). In other words, under a combined wetting and warming scenario, peak discharge responds more strongly to changes in precipitation (rainfall) in the hemiboreal baseline climate, while peak discharge is comparatively more sensitive to warming in the boreal baseline climate, due to snowpack depletion effects.

Fig. 11c reveals that basin landscape can lead to asymmetric changes in peak Q in response to climate change. Despite an increase in available mean effective precipitation in response to combined warming and wetting in Montmorency under its own (boreal) climate (Fig. 8c), the mean runoff declines (Supplementary Fig. S6c), which in turn leads to a 1% decline in peak Q (Fig. 11c). The same degree of warming and wetting also leads to a similar decrease in peak Q (by 0.5%) in the artificially deforested Montmorency. These findings reveal that the soil characteristics play the major role in governing the response of peak flow to climate change and that the forested porous soils buffer the increased mean water fluxes and lead to decreased mean runoff (Supplementary Fig. S6c) and peak Q (Fig. 11c). This effect is well illustrated in Fig. 9, which shows that the buffering capacity of the Montmorency-type landscape increases under a warming/wetting scenario, especially

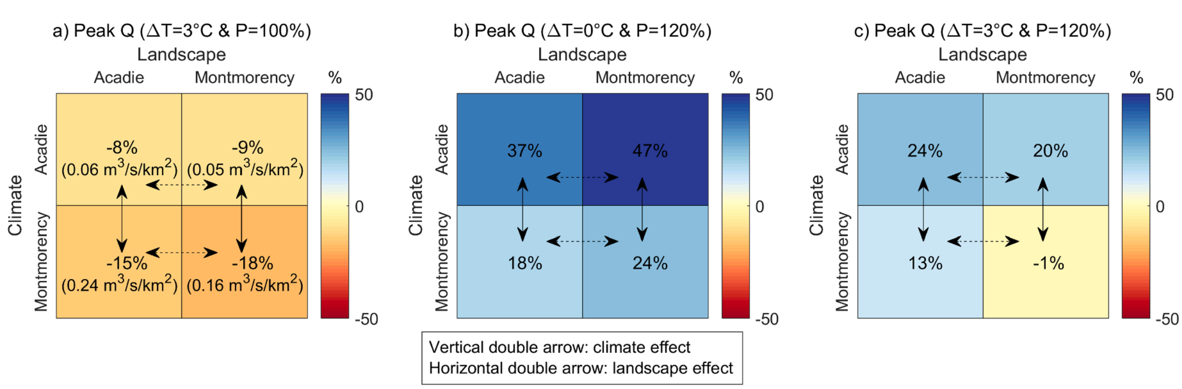

Fig. 11. Influence of basin landscape and climate conditions on peak streamflow from permuted baseline climate experiments. (a) 3 °C warming; (b) 20% wetting; (c) 3 °C warming with 20% wetting. Peak streamflow under baseline climate conditions is displayed in parenthesis in panel a.

for the boreal baseline climate (Fig. 9b, c).

Contrary to the Montmorency-type landscape under boreal climate, the Acadie-type landscape under boreal climate shows an increase in peak Q in response to warming and wetting (Fig. 11c). This suggests that the increase in mean effective precipitation simulated for Acadie under the boreal baseline climate (Fig. 8c) translates into an increase in mean runoff (Fig. S6c), and as a result peak Q increases (Fig. 11c). Hence our results suggest that the reduced infiltration and storage capacity of the agricultural soils are less apt to buffer the increased runoff and lead to increased peak flow.

4. Discussion

4.1. Simulated water balance in Montmorency

Our model simulations demonstrate that 28% of the total precipitation left as evapotranspiration and sublimation in Montmorency River Basin for the 2005–2019 period, which broadly agrees with observations reported for the BEREV subbasin of Montmorency (34%: Isabelle et al., 2020; 30%: Schilling et al., 2021). Simulated sublimation itself only accounted for about 10% of the annual precipitation in the basin, which is within the range (5%–13%) reported by Parajuli et al. (2020) for the BEREV subbasin of Montmorency.

Over the 2005–2019 period, snowmelt contributed 46% of the simulated annual streamflow in the Montmorency River Basin, which is in line with the contribution (45%–47%) reported by Isabelle et al. (2020) for the BEREV subbasin of Montmorency over the 2016–2018 period. The simulated soil moisture also exhibited a good match with the observations (Figs. S3 and S4), suggesting that the model represents fairly well infiltration and thus the recharge process. Our results highlight that there is a significant contribution from groundwater flow to streamflow as baseflow in Montmorency. Groundwater flow provided 48% of the annual streamflow for the 14 years of simulation period in Montmorency. This is of a similar magnitude with the 50% groundwater contribution to streamflow calculated from on-site tracer experiments in the BEREV subbasin of Montmorency for the 2017–2018 period (Schilling et al., 2021). Despite the relatively simple treatment of groundwater flow in CHRM, these results suggest an adequate representation of groundwater contribution to streamflow.

4.2. Landscape and climate influence on present-day hydrological processes

When both basins are forced by the snowier boreal climate, less snow accumulates in the Montmorency-type landscape (Fig. 7a) due to higher (canopy) sublimation losses (Fig. 3c), which is consistent with previous studies that reported less snow accumulation in a forested basin than in an agricultural basin in Russia (Gelfan et al., 2004), and less snow accumulation in forested sites compared with nearby open environments in cold regions of Europe (Koivusalo and Kokkonen, 2002) and North America (Jost et al., 2007; Varhola et al., 2010). On the other hand, when forced by the hemiboreal climate, reduced mid-winter snowmelt was simulated in the forested basin (due to shading) which compensated the canopy interception losses, such that both basins exhibited similar maximum snow accumulations (Fig. 7a). This is in agreement with Varhola et al. (2010) and Veatch et al. (2009) who have shown that reduced mid-winter and spring ablation rates in forests can offset the lesser accumulation, resulting in snowpacks being thicker in forests than in open areas (Gelfan et al., 2004; Veatch et al., 2009).

Although higher than in Acadie (Fig. 3d), the ratio of canopy interception sublimation to snowfall (21%) in Montmorency (Fig. 3c) is lower than the 25–45% values reported from colder and drier boreal forest environments (Essery et al., 2003) such as the southern boreal forest in Saskatchewan (Pomeroy and Gray, 1995; Pomeroy et al., 1998). This is because Montmorency River Basin has a humid climate (Dfc; Peel et al., 2007) which limits the sublimation (Essery and Pomeroy, 2001). However, it is important to note that although the canopy sublimation ratio is lower for Montmorency than for the drier climate of Saskatchewan, the absolute amount is greater, which is explained by the higher amount of snowfall (Fig. 3a) and hence longer duration for which intercepted snow is exposed to the atmosphere in Montmorency. On the other hand, the simulated canopy sublimation to snowfall ratio in Montmorency (Fig. 3c) is higher than the 10% canopy sublimation loss estimated for the Umpqua National Forest in Oregon (US) (Storck et al., 2002) where the climate is warmer and more humid, which results in more rapid unloading and in limited sublimation (Essery et al., 2003). The canopy sublimation losses in Montmorency decrease by about 7% per °C (Fig. 3c), which is similar to that recently reported for a southern boreal forest of western Canada (6% per °C) (He et al., 2021). Decreasing canopy sublimation losses in response to warming was also found to explain the increase in effective precipitation in the forested landscape (Fig. 8).

When forced by common climate conditions, the annual peak discharge was found to be higher in the Acadie-type landscape (Fig. 11a), which shows that the lower buffering capacity (i.e., reduced infiltration and storage capacity) (Fig. 9a) of agricultural soils favor runoff, despite the lower slopes compared to the more rugged Montmorency-type landscape.

4.3. Landscape and climate influence on hydrological sensitivity to climate change

Slower snowmelt rates were simulated in both basins in response to warming (Fig. 4c and f), which is consistent with findings in some other snowmelt dominated regions located in Europe (López-Moreno et al., 2013) and North America (Musselman et al., 2017; Rasouli et al., 2014; Rasouli et al., 2019b; Trujillo and Molotch, 2014) where snowmelt occurs before the summer solstice (Essery et al., 2020). In contrast, faster snowmelt rates could occur in the future when limited warming (<1°C for Acadie, and <2.5 °C for Montmorency) is accompanied by increasing precipitation (Fig. 4c, f), which is similar to the increased snowmelt rates in response to warming reported for relatively deeper SWE regions (>150 mm) across the Northern Hemisphere (Wu et al., 2018) and increased rates in the northern boreal forest of the Northwest Territories, Canada where snowfall increases along with rising temperature (Krogh and Pomeroy, 2019). These results suggest that changes in snowmelt rate are influenced not only by warming but also by increasing precipitation, and future snowmelt rate can be either lower or higher depending on the regional climate and snow regimes and whether the snowmelt period is shifted by changing climate into a lower or higher insolation season.

Similar to the findings from the inter-comparison study of Carey et al. (2010) in northern basins, the boreal climate with higher precipitation (as snow) exhibited a lower correlation between precipitation and discharge than the hemiboreal climate (Fig. 9). The declining snow accumulation in response to warming, therefore, leads to more synchronized precipitation and discharge. Regarding the impact of basin landscape, the increase in correlation between P and Q in response to climate change is greater in the Acadie-type landscape which presents more compacted soils with smaller storage capacity (smaller buffering capacity; Fig. 9a), highlighting the influence of soil characteristics on the climate sensitivity of precipitation-discharge synchronicity.

Streamflow regimes showed less sensitivity to warming (Fig. 11a) compared to the changes in peak snowpack accumulation (Fig. 7a), under all baseline climate and landscape combinations. This finding is in agreement with the conclusion from a recent worldwide comparison of high mountain snowpack and streamflow sensitivities to climate warming (López-Moreno et al., 2020). Notwithstanding, the colder boreal climate was found to enhance the sensitivity of peak Q to warming due to the persisting dominant influence of the snow cycle on streamflow, but to reduce the sensitivity to wetting, compared to the hemiboreal climate (Fig. 11). In other words, a higher snowfall fraction in a basin (Fig. 3a) buffers against extreme streamflow response to increasing precipitation, which is consistent with the conclusions of previous studies conducted in North America by Sankarasubramanian et al. (2001), Dethier et al. (2020) and more recently by Robles et al. (2021). In both Montmorency and Acadie, the streamflow regimes were projected to shift towards a more rainfall dominated regime under warmer temperatures, with faster transition projected for Acadie (Fig. 5e).

Peak discharge in basins with a mild, hemiboreal winter climate could increase in response to a combination of 3 °C warming and 20% increasing precipitation due the substantial conversion of snowfall to rainfall combined with enhanced winter snowmelt, which together lead to higher surface runoff extremes in winter (Fig. 11c). The Acadie-type landscape, with its reduced buffering capacity (Fig. 9c), is more sensitive to this scenario than the Montmorency-type landscape, whose increased buffering capacity (Fig. 9c) attenuates extreme rainfall-snowmelt events. This finding is in agreement with a recent study over selected North American mountain basins where higher storage and infiltration capacity (due to forestation) resulted in decreased high flows in a warmer and wetter future (Rasouli et al., 2019a).

Conversely, under a colder boreal climate and for the same climate change scenario (3 °C warming and 20% increasing precipitation), the streamflow remains largely synchronized with the snowpack regime. The bulk of snowmelt continues to occur in the spring with more limited conversion of snowfall to rainfall, which attenuates extreme runoff events. However, basins respond differently to the increased water inputs: in the more impervious Acadie-type landscape, the amount of runoff (but not its intensity) increases and leads to higher peak discharge in the spring, while the porous soils of Montmorency-type landscape largely buffer the increased flux, resulting in decreased runoff amount and lower peak discharge (Fig. 11c).

4.4. Limitations and prospects

This study aimed to assess how present-day climate and landscape conditions modulate the hydrological sensitivity, while the potential impacts of climate change on vegetation such as the reduction in stomatal conductance (Le et al., 2011), decline in canopy LAI caused by warming (Boulanger et al., 2017) and the forest disturbance due to fire or insects (Seidl et al., 2017) were not considered. This imposes a limitation on the computed hydrological sensitivities since including future vegetation changes is likely to modulate the hydrological responses to climate change in snowmelt dominated basins (Rasouli et al., 2019a) and elsewhere (Gerten et al., 2004). However, including feedbacks from changing vegetation due to climate change would complicate isolating the effect of current vegetation on the hydrological sensitivity to climate change. This study thus demonstrates how current landscape conditions affect the hydrological sensitivity, but future changes in vegetation and/or land use could invariably contribute to changes in water fluxes within the catchments and should be considered for more accurate climate change impact assessments.

It is also worth noting that many of the subjective choices made when choosing parameterizations and assigning parameters in physically-based hydrological models can have a significant impact on the magnitude of the output uncertainty. The separation of precipitation into rainfall or snowfall is one of the most sensitive parameterizations when simulating cold regions hydrological processes (Harder and Pomeroy, 2013). Underestimation (overestimation) of rainfall can advect less (more) energy to snowpack, decrease (increase) snowpack liquid content, lead to earlier (later) warming and ripening and in turn impact the magnitude and timing of snowmelt streamflow peak flow (Pomeroy et al, 2016). In this study, the total precipitation was partitioned between liquid and solid precipitation using a psychometric energy balance method proposed by Harder and Pomeroy (2013). Rainfall fraction was calculated as a function of the temperature of falling precipitation estimated using the air temperature and relative humidity. Future studies could use in situ records of precipitation phase made in the BEREV watershed (Pierre et al., 2019) to calibrate air temperature-precipitation phase relationships to reduce errors in phase partitioning. While most of the model parameters were prescribed based on observations and previous studies within the basins, some parameters such as maximum canopy snow load capacity and maximum active layer thickness of snow were borrowed from basins with similar landscape conditions, which introduce uncertainties in the hydrological models. Future studies are required to perform detailed parameter sensitivity analyses to fully quantify the uncertainty in hydrological simulations.

5. Conclusion

A physically based hydrological model was set up using the Cold Regions Hydrological Modelling (CRHM) platform, driven with current climate data and perturbed by projected annually uniform changes in temperature and precipitation to analyze the influences of these changes on the hydrological regime of two archetypical climates and landscapes of the St-Lawrence River in eastern Canada, namely a rugged and forested basin with boreal climate (Montmorency, north shore) and a relatively level agricultural basin with hemiboreal climate (Acadie, south shore).

On a regional perspective, the main finding is that although these basins are not very far from each other (<400 km), their distinct present-day meteorological conditions lead to different hydrological sensitivities to the same climate perturbation. The faster and more dramatic hydrological changes were simulated for Acadie in response to warming, due to its milder present-day climate (Table 1). By the mid-century, the snow accumulation in Acadie is projected to almost completely (by 80%) vanish, while the peak snow accumulation in Montmorency declines by less than 50%. By the end of the century, Acadie is projected to be completely dominated by rainfall-runoff mechanisms, whereas the snowmelt continues to play a dominant role on the peak runoff in Montmorency. These results have important implications for regional water management adaptation strategies.

In addition to the regional findings, this study more broadly illustrates the contrasted hydrological behaviours that can arise under present and future climates in cold forested and agricultural landscapes of the globe. Competing mechanisms were found between the current climate conditions and the landscape characteristics, which dictate the amounts of snow accumulations. This interplay between regional climate and the landscape features greatly impacts the climate sensitivity of the basins, i.e., how snow accumulation, snowmelt, runoff and streamflow responded to a common climate change signal. The major findings are as follows:

i) A forested landscape can present a lower or equal amount of snow accumulation compared to an agricultural environment, depending on the regional climate that governs the compensation effect between canopy sublimation losses on the one hand, and reduced snowmelt due to forest shading on the other hand.
ii) The response of peak SWE to climate change depends only on regional climate and is not impacted by basin landscape.
iii) Sublimation rate declines in a warmer climate due to faster decreasing snow availability.
iv) Basin topography and soil characteristics influence in different ways the synchronization between precipitation and discharge. Steeper basin slopes favor a tighter coupling between precipitation and discharge, whereas the more porous soils with higher infiltration capacity dampen the sensitivity of this coupling to climate.
v) A more impervious agricultural landscape tends to translate the climate variability to higher peak flows, whereas a forested landscape can “absorb” the same degree of variability and even lead to reduced flows. Afforestation, therefore, can be an

important way of mitigating the future climate change-induced hydrological extremes, particularly in colder cold regions.

vi) The attenuation of extreme flows by forested landscapes in response to climate change is climate-dependent: it is more efficient in present-day colder temperature regime than in milder cold regions, where greater conversions from snowfall to rainfall in response to warming can overwhelm the buffering capacity of even the deep forested soils.

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CRediT authorship contribution statement

Okan Aygün: Methodology, Formal analysis, Data curation, Writing – original draft. **Christophe Kinnard:** Conceptualization, Methodology, Formal analysis, Writing – review & editing, Supervision, Project administration, Funding acquisition. **Stéphane Campeau:** Conceptualization, Writing – review & editing, Supervision, Funding acquisition. **John W. Pomeroy:** Writing – review & editing.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

Data will be made available on request.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.jhydrol.2022.128595.

References

Annandale, J., Jovanovic, N., Benade, N., Allen, R., 2002. Software for missing data error analysis of Penman-Monteith reference evapotranspiration. Irrig. Sci. 21 (2), 57–67. https://doi.org/10.1007/s002710100047.

Arheimer, B., Donnelly, C., Strömqvist, J., 2013. Large-scale effects of climate change on water resources in Sweden and Europe. VATTEN: J. Water Manage. Res. 69, 201–207.

Arnell, N.W., Gosling, S.N., 2016. The impacts of climate change on river flood risk at the global scale. Clim. Change 134 (3), 387–401.

Aygün, O., Kinnard, C., Campeau, S., 2020a. Impacts of climate change on the hydrology of northern midlatitude cold regions. Prog. Phys. Geogr.: Earth Environ. 44 (3), 338–375. https://doi.org/10.1177/0309133319878123.

Aygün, O., Kinnard, C., Campeau, S., Krogh, S.A., 2020b. Shifting hydrological processes in a Canadian agroforested catchment due to a warmer and wetter climate. Water 12 (3), 739. https://doi.org/10.3390/w12030739.

Aygün, O., Kinnard, C., Campeau, S., 2021. Responses of soil erosion to warming and wetting in a cold Canadian agricultural catchment. CATENA 201, 105184.

Barry, R., Plamondon, A.P., Stein, J., 1988. Hydrologic soil properties and application of a soil moisture model in a balsam fir forest. Can. J. For. Res. 18 (4), 427–434.

Barry, R., Prévost, M., Stein, J., Plamondon, A.P., 1990. Simulation of snowmelt runoff pathways on the Lac Laflamme watershed. J. Hydrol. 113 (1–4), 103–121.

Bergeron, O., 2016. Guide d'utilisation 2016 – Grilles climatiques quotidiennes du Programme de surveillance du climat du Québec. Ministère du Développement Durable de l'Environnement et de la Lutte contre les Changements Climatiques du Québec, 33 pp.

Berghuijs, W., Woods, R., Hrachowitz, M., 2014. A precipitation shift from snow towards rain leads to a decrease in streamflow. Nat. Clim. Change 4 (7), 583–586.

Boulanger, Y., et al., 2017. Climate change impacts on forest landscapes along the Canadian southern boreal forest transition zone. Landscape Ecol. 32 (7), 1415–1431.

Brooks, R., Corey, A., 1964. Hydraulic Properties of Porous Media, Hydrology Papers no. 3. Colorado State University, ft. Hydrology Papers (Colorado State University) (3). , 37.

Brown, R.D., Mote, P.W., 2009. The response of Northern Hemisphere snow cover to a changing climate. J. Clim. 22 (8), 2124–2145.

Brown, A.E., Zhang, L., McMahon, T.A., Western, A.W., Vertessy, R.A., 2005. A review of paired catchment studies for determining changes in water yield resulting from alterations in vegetation. J. Hydrol. 310 (1–4), 28–61. https://doi.org/10.1016/j.jhydrol.2004.12.010.

Broxton, P., et al., 2015. Quantifying the effects of vegetation structure on snow accumulation and ablation in mixed-conifer forests. Ecohydrology 8 (6), 1073–1094.

Buttle, J., Creed, I., Moore, R., 2009. Advances in Canadian forest hydrology, 2003–2007. Can. Water Resour. J. 34 (2), 113–126.

Carey, S.K., et al., 2010. Inter-comparison of hydro-climatic regimes across northern catchments: synchronicity, resistance and resilience. Hydrol. Process. 24 (24), 3591–3602.

Chandler, D.G., 2006. Reversibility of forest conversion impacts on water budgets in tropical karst terrain. For. Ecol. Manage. 224 (1–2), 95–103. https://doi.org/10.1016/j.foreco.2005.12.010.

Charron, I., 2016. A Guidebook on Climate Scenarios: Using Climate Information to Guide Adaptation Research and Decisions. Ouranos, Montreal, Canada.

Dethier, E.N., Sartain, S.L., Renshaw, C.E., Magilligan, F.J., 2020. Spatially coherent regional changes in seasonal extreme streamflow events in the United States and Canada since 1950. Sci. Adv. 6 (49), eaba5939.

Dias, L.C.P., Macedo, M.N., Costa, M.H., Coe, M.T., Neill, C., 2015. Effects of land cover change on evapotranspiration and streamflow of small catchments in the Upper Xingu River Basin, Central Brazil. J. Hydrol.: Reg. Stud. 4, 108–122. https://doi.org/10.1016/j.ejrh.2015.05.010.

D'Orangeville, L., Houle, D., Duchesne, L., Côté, B., 2016. Can the Canadian drought code predict low soil moisture anomalies in the mineral soil? An analysis of 15 years of soil moisture data from three forest ecosystems in Eastern Canada. Ecohydrology 9 (2), 238–247.

Dornes, P.F., Pomeroy, J.W., Pietroniro, A., Carey, S.K., Quinton, W.L., 2008. Influence of landscape aggregation in modelling snow-cover ablation and snowmelt runoff in a sub-arctic mountainous environment. Hydrol. Sci. J. 53 (4), 725–740.

Easton, Z.M., Gérard-Marchant, P., Walter, M.T., Petrovic, A.M., Steenhuis, T.S., 2007. Hydrologic assessment of an urban variable source watershed in the northeast United States. Water Resour. Res. 43 (3) https://doi.org/10.1029/2006WR005076.

Essery, R., et al., 2020. Snow cover duration trends observed at sites and predicted by multiple models. Cryosphere 14 (12), 4687–4698.

Essery, R., Pomeroy, J., 2001. In: Sublimation of snow intercepted by coniferous forest canopies in a climate model. 6th IAHS Scientific Assembly IAHS PUBLICATION, Maastricht, Netherlands, pp. 343–348.

Essery, R., Pomeroy, J., Parviainen, J., Storck, P., 2003. Sublimation of snow from coniferous forests in a climate model. J. Clim. 16 (11), 1855–1864.

Fang, X., et al., 2013. Multi-variable evaluation of hydrological model predictions for a headwater basin in the Canadian Rocky Mountains. Hydrol. Earth Syst. Sci. 17 (4), 1635–1659. https://doi.org/10.5194/hess-17-1635-2013.

Fang, X., Pomeroy, J.W., Westbrook, C.J., Guo, X., Minke, A.G., Brown, T., 2010. Prediction of snowmelt derived streamflow in a wetland dominated prairie basin. Hydrol. Earth Syst. Sci. 14 (6), 991–1006.

Gelfan, A., Pomeroy, J., Kuchment, L., 2004. Modeling forest cover influences on snow accumulation, sublimation, and melt. J. Hydrometeorol. 5 (5), 785–803.

Germer, S., Neill, C., Krusche, A.V., Elsenbeer, H., 2010. Influence of land-use change on near-surface hydrological processes: undisturbed forest to pasture. J. Hydrol. 380 (3–4), 473–480. https://doi.org/10.1016/j.jhydrol.2009.11.022.

Gerten, D., Schaphoff, S., Haberlandt, U., Lucht, W., Sitch, S., 2004. Terrestrial vegetation and water balance—hydrological evaluation of a dynamic global vegetation model. J. Hydrol. 286 (1–4), 249–270.

Guillemette, F., Plamondon, A.P., Prévost, M., Lévesque, D., 2005. Rainfall generated stormflow response to clearcutting a boreal forest: peak flow comparison with 50 world-wide basin studies. J. Hydrol. 946 302(1–4), 137–153.

Gupta, H.V., Kling, H., Yilmaz, K.K., Martinez, G.F., 2009. Decomposition of the mean squared error and NSE performance criteria: implications for improving hydrological modelling. J. Hydrol. 377 (1–2), 80–91.

Hadiwijaya, B., Pepin, S., Isabelle, P.E., Nadeau, D.F., 2020. The dynamics of transpiration to evapotranspiration ratio under wet and dry canopy conditions in a humid boreal forest. Forests 11 (2), 237.

Harder, Phillip, Pomeroy, John, 2013. Estimating precipitation phase using a psychrometric energy balance method. Hydrological Processes. https://doi.org/10.1002/hyp.9799.

He, Z., Pomeroy, J., Fang, X., Peterson, A., 2021. Sensitivity analysis of hydrological processes to perturbed climate in a southern boreal forest basin. J. Hydrol. 601, 126706.
Hedstrom, N., Pomeroy, J., 1998. Measurements and modelling of snow interception in the boreal forest. Hydrol. Process. 12 (10–11), 1611–1625.
Herron, N., Wilson, C., 2001. A water balance approach to assessing the hydrologic buffering potential of an alluvial fan. Water Resour. Res. 37 (2), 341–351.
Houle, D., Bouffard, A., Duchesne, L., Logan, T., Harvey, R., 2012. Projections of future soil temperature and water content for three southern Quebec forested sites. J. Clim. 25 (21), 7690–7701.
Houle, D., Lajoie, G., Duchesne, L., 2016. Major losses of nutrients following a severe drought in a boreal forest. Nat. Plants 2 (12), 1–5.
Isabelle, P.E., Nadeau, D.F., Anctil, F., Rousseau, A.N., Jutras, S., Music, B., 2020. Impacts of high precipitation on the energy and water budgets of a humid boreal forest. Agric. For. Meteorol. 280, 107813.
Jost, G., Weiler, M., Gluns, D.R., Alila, Y., 2007. The influence of forest and topography on snow accumulation and melt at the watershed-scale. J. Hydrol. 347 (1–2), 101–115. https://doi.org/10.1016/j.jhydrol.2007.09.006.
Jutras, M.-F., 2012. Modeling Stream Discharge in Forest Catchments across Canada: Hydraulic Conductivity Calibrations. University of New Brunswick, New Brunswick, Canada.
Kay, A., Rudd, A., Fry, M., Nash, G., Allen, S., 2021. Climate change impacts on peak river flows: combining national-scale hydrological modelling and probabilistic projections. Clim. Risk Manage. 31, 100263.
Kellomäki, S., Maajärvi, M., Strandman, H., Kilpeläinen, A., Peltola, H., 2010. Model computations on the climate change effects on snow cover, soil moisture and soil frost in the boreal conditions over Finland. Silva Fennica 44 (2), 213–233.
Koivusalo, H., Kokkonen, T., 2002. Snow processes in a forest clearing and in a coniferous forest. J. Hydrol. 262 (1–4), 145–164. https://doi.org/10.1016/S0022-1694(02)00031-8.
Köplin, N., Schädler, B., Viviroli, D., Weingartner, R., 2012. Relating climate change signals and physiographic catchment properties to clustered hydrological response types. Hydrol. Earth Syst. Sci. 16 (7), 2267–2283.
Krogh, S.A., Pomeroy, J.W., 2019. Impact of future climate and vegetation on the hydrology of an arctic headwater basin at the Tundra-Taiga transition. J. Hydrometeorol. 20 (2), 197–215. https://doi.org/10.1175/JHM-D-18-0187.1.
Kuentz, A., Arheimer, B., Hundecha, Y., Wagener, T., 2017. Understanding hydrologic variability across Europe through catchment classification. Hydrol. Earth Syst. Sci. 21 (6), 2863–2879.
Lavigne, M.P., 2007. Modélisation du régime hydrologique et de l'impact des coupes forestières sur l'écoulement du ruisseau des Eaux-Volées à l'aide d'Hydrotel (Doctoral Dissertation). Université du Québec, Institut national de la recherche scientifique.
Le, P.V., Kumar, P., Drewry, D.T., 2011. Implications for the hydrologic cycle under climate change due to the expansion of bioenergy crops in the Midwestern United States. Proc. Natl. Acad. Sci. U.S.A. 108 (37), 15085–15090.
Li, L., Pomeroy, J.W., 1997. Estimates of threshold wind speeds for snow transport using meteorological data. J. Appl. Meteorol. 36 (3), 205–213.
López-Moreno, J., Pomeroy, J., Revuelto, J., Vicente-Serrano, S., 2013. Response of snow processes to climate change: spatial variability in a small basin in the Spanish Pyrenees. Hydrol. Process. 27 (18), 2637–2650. https://doi.org/10.1002/hyp.9408.
López-Moreno, J., Pomeroy, J., Alonso-González, E., Morán-Tejeda, E., Revuelto, J., 2020. Decoupling of warming mountain snowpacks from hydrological regimes. Environ. Res. Lett. 15 (11), 114006.
Minville, M., Brissette, F., Leconte, R., 2008. Uncertainty of the impact of climate change on the hydrology of a nordic watershed. J. Hydrol. 358 (1–2), 70–83. https://doi.org/10.1016/j.jhydrol.2008.05.033.
Moore, R., Wondzell, S., 2005. Physical hydrology and the effects of forest harvesting in the Pacific Northwest: a review 1. JAWRA: J. Am. Water Resour. Assoc. 41 (4), 763–784.
Musselman, K.N., Clark, M.P., Liu, C., Ikeda, K., Rasmussen, R., 2017. Slower snowmelt in a warmer world. Nat. Clim. Change 7 (3), 214–219. https://doi.org/10.1038/nclimate3225.
Ouimet, R., Duchesne, L., 2005. Base cation mineral weathering and total release rates from soils in three calibrated forest watersheds on the Canadian Boreal Shield. Can. J. Soil Sci. 85 (2), 245–260. https://doi.org/10.4141/S04-061.
Padrón, R.S., Gudmundsson, L., Greve, P., Seneviratne, S.I., 2017. Large-scale controls of the surface water balance over land: insights from a systematic review and meta-analysis. Water Resour. Res. 53 (11), 9659–9678.
Parajuli, A., Nadeau, D.F., Anctil, F., Schilling, O.S., Jutras, S., 2020. Does data availability constrain temperature-index snow models? A case study in a humid boreal forest. Water 12 (8), 2284.
Peel, M.C., Finlayson, B.L., McMahon, T.A., 2007. Updated world map of the Köppen-Geiger climate classification. Hydrol. Earth Syst. Sci. 11, 1633–1644. https://doi.org/10.5194/hess-11-1633-2007.
Pierre, A., Jutras, S., Smith, C., Kochendorfer, J., Fortin, V., Anctil, F., 2019. Evaluation of catch efficiency transfer functions for unshielded and single-alter-shielded solid precipitation measurements. J. Atmos. Oceanic Technol. 36 (5), 865–881.
Plamondon, A.P., Prévost, M., Naud, C., R., 1984. Accumulation et fonte de la neige en milieux boisé et déboisé. Géog. Phys. Quatern. 38 (1), 27–35.
Pomeroy, J., et al., 2007. The cold regions hydrological model: a platform for basing process representation and model structure on physical evidence. Hydrol. Process. 21 (19), 2650–2667.
Pomeroy, J., Fang, X., Ellis, C., 2012. Sensitivity of snowmelt hydrology in Marmot Creek, Alberta, to forest cover disturbance. Hydrol. Process. 26 (12), 1891–1904. https://doi.org/10.1002/hyp.9248.
Pomeroy, J., Fang, X., Marks, D.G., 2016. The cold rain-on-snow event of June 2013 in the Canadian Rockies-Characteristics and diagnosis. Hydrol. Process. 30 (17), 2899–2914. https://doi.org/10.1002/hyp.10905.
Pomeroy, J., Gray, D., 1995. Snowcover accumulation, relocation and management. Bull. Int. Soc. Soil Sci. 88 (2).
Pomeroy, J., Parviainen, J., Hedstrom, N., Gray, D., 1998. Coupled modelling of forest snow interception and sublimation. Hydrol. Process. 12 (15), 2317–2337.
Prévost, M., Barry, R., Stein, J., Plamondon, A.P., 1991. Snowmelt modeling in a balsam fir forest: comparison between an energy balance model and other simplified models. Can. J. For. Res. 21 (1), 1–10.
Prudhomme, C., Wilby, R.L., Crooks, S., Kay, A.L., Reynard, N.S., 2010. Scenario-neutral approach to climate change impact studies: application to flood risk. J. Hydrol. 390 (3–4), 198–209.
Prudhomme, C., Crooks, S., Kay, A.L., Reynard, N., 2013. Climate change and river flooding: part 1 classifying the sensitivity of British catchments. Clim. Change 119 (3–4), 933–948.
Räisänen, J., Eklund, J., 2012. 21st century changes in snow climate in Northern Europe: a high-resolution view from ENSEMBLES regional climate models. Clim. Dyn. 38 (11–12), 2575–2591.
Rasouli, K., Pomeroy, J.W., Janowicz, J.R., Carey, S.K., Williams, T.J., 2014. Hydrological sensitivity of a northern mountain basin to climate change. Hydrol. Process. 28 (14), 4191–4208. https://doi.org/10.1002/hyp.10244.
Rasouli, K., Pomeroy, J.W., Marks, D.G., 2015. Snowpack sensitivity to perturbed climate in a cool mid-latitude mountain catchment. Hydrol. Process. 29 (18), 3925–3940. https://doi.org/10.1002/hyp.10587.
Rasouli, K., Pomeroy, J.W., Whitfield, P.H., 2019a. Are the effects of vegetation and soil changes as important as climate change impacts on hydrological processes? Hydrol. Earth Syst. Sci. 23 (12), 4933–4954.
Rasouli, K., Pomeroy, J.W., Whitfield, P.H., 2019b. Hydrological responses of headwater basins to monthly perturbed climate in the North American cordillera. J. Hydrometeorol. 20 (5), 863–882.
Rasouli, K., Pomeroy, J.W., Whitfield, P.H., 2022. The sensitivity of snow hydrology to changes in air temperature and precipitation in three North American headwater basins. J. Hydrol. 127460.
Rice, J.S., Emanuel, R.E., Vose, J.M., Nelson, S.A., 2015. Continental US streamflow trends from 1940 to 2009 and their relationships with watershed spatial characteristics. Water Resour. Res. 51 (8), 6262–6275.
Robinet, J., et al., 2018. Impacts of forest conversion and agriculture practices on water pathways in Southern Brazil. Hydrol. Process. 32 (15), 2304–2317. https://doi.org/10.1002/hyp.13155.
Robles, M.D., Hammond, J.C., Kampf, S.K., Biederman, J.A., Demaria, E., 2021. Winter inputs buffer streamflow sensitivity to snowpack losses in the salt river watershed in the lower Colorado river basin. Water 13 (1), 3.
Sankarasubramanian, A., Vogel, R.M., Limbrunner, J.F., 2001. Climate elasticity of streamflow in the United States. Water Resour. Res. 37 (6), 1771–1781.
Savary, S., Rousseau, A.N., Quilbé, R., 2009. Assessing the effects of historical land cover changes on runoff and low flows using remote sensing and hydrological modeling. J. Hydrol. Eng. 14 (6), 575–587. https://doi.org/10.1061/(ASCE)HE.1943-5584.0000024.
Sawicz, K., et al., 2014. Characterizing hydrologic change through catchment classification. Hydrol. Earth Syst. Sci. 18 (1), 273.
Schilling, O.S., et al., 2021. Quantifying groundwater recharge dynamics and unsaturated zone processes in snow-dominated catchments via on-site dissolved gas analysis. Water Resour. Res. 57 (2) e2020WR028479.
Schnorbus, M., Werner, A., Bennett, K., 2014. Impacts of climate change in three hydrologic regimes in British Columbia, Canada. Hydrol. Processes 28 (3), 1170–1189.
Seidl, R., et al., 2017. Forest disturbances under climate change. Nat. Clim. Change 7 (6), 395–402.
Smith, C.D., 2007. Correcting the wind bias in snowfall measurements made with a Geonor T-200B precipitation gauge and alter wind shield. Proceedings of the 14th SMOI, San Antonio.
Stonevičius, E., Rimkus, E., Štaras, A., Kažys, J., Valiuškevičius, G., 2017. Climate change impact on the Nemunas River basin hydrology in the 21st century. Boreal Environ. Res. 22 (1), 49–65.
Storck, P., Lettenmaier, D.P., Bolton, S.M., 2002. Measurement of snow interception and canopy effects on snow accumulation and melt in a mountainous maritime climate, Oregon, United States. Water Resour. Res. 38 (11) https://doi.org/10.1029/2002WR001281.
Sulis, M., Paniconi, C., Marrocu, M., Huard, D., Chaumont, D., 2012. Hydrologic response to multimodel climate output using a physically based model of groundwater/surface water interactions. Water Resour. Res. 48 (12).
Teutschbein, C., Grabs, T., Karlsen, R.H., Laudon, H., Bishop, K., 2015. Hydrological response to changing climate conditions: spatial streamflow variability in the boreal region. Water Resour. Res. 51 (12), 9425–9446.
Tremblay, Y., Rousseau, A.N., Plamondon, A.P., Lévesque, D., Jutras, S., 2008. Rainfall peak flow response to clearcutting 50% of three small watersheds in a boreal forest, Montmorency Forest, Québec. J. Hydrol. 352 (1–2), 67–76.
Tremblay, Y., Rousseau, A.N., Plamondon, A.P., Lévesque, D., Prévost, M., 2009. Changes in stream water quality due to logging of the boreal forest in the Montmorency Forest, Québec. Hydrol. Processes: Int. J. 23 (5), 764–776.
Trujillo, E., Molotch, N.P., 2014. Snowpack regimes of the western United States. Water Resour. Res. 50 (7), 5611–5623. https://doi.org/10.1002/2013WR014753.
Van Tiel, M., Van Loon, A.F., Seibert, J., Stahl, K., 2021. Hydrological response to warm and dry weather: do glaciers compensate? Hydrol. Earth Syst. Sci. 25 (6), 3245–3265.

Varhola, A., Coops, N.C., Weiler, M., Moore, R.D., 2010. Forest canopy effects on snow accumulation and ablation: an integrative review of empirical results. J. Hydrol. 392 (3–4), 219–233. https://doi.org/10.1016/j.jhydrol.2010.08.009.

Veatch, W., Brooks, P., Gustafson, J., Molotch, N., 2009. Quantifying the effects of forest canopy cover on net snow accumulation at a continental, mid-latitude site. Ecohydrology: ecosystems, land and water process interactions. Ecohydrogeomorphology 2 (2), 115–128. https://doi.org/10.1002/eco.45.

Wetterhall, F., Graham, L., Andreasson, J., Rosberg, J., Yang, W., 2011. Using ensemble climate projections to assess probabilistic hydrological change in the Nordic region. Nat. Hazards Earth Syst. Sci. 11 (8), 2295–2306.

Whateley, S., Steinschneider, S., Brown, C., 2014. A climate change range-based method for estimating robustness for water resources supply. Water Resour. Res. 50 (11), 8944–8961.

Whitaker, A., Alila, Y., Beckers, J., Toews, D., 2002. Evaluating peak flow sensitivity to clear-cutting in different elevation bands of a snowmelt-dominated mountainous catchment. Water Resour. Res. 38 (9), 11-1-11-17.

Williams, C.A., et al., 2012. Climate and vegetation controls on the surface water balance: synthesis of evapotranspiration measured across a global network of flux towers. Water Resour. Res. 48 (6).

Wu, X., Che, T., Li, X., Wang, N., Yang, X., 2018. Slower snowmelt in spring along with climate warming across the Northern Hemisphere. Geophys. Res. Lett. 45 (22), 12331–12339.

Zhou, J., et al., 2014. Simulating cold regions hydrological processes using a modular model in the west of China. J. Hydrol. 509, 13–24. https://doi.org/10.1016/j.jhydrol.2013.11.013.

Ziegler, A.D., et al., 2004. Hydrological consequences of landscape fragmentation in mountainous northern Vietnam: evidence of accelerated overland flow generation. J. Hydrol. 287 (1–4), 124–146. https://doi.org/10.1016/j.jhydrol.2003.09.027.