UNIVERSITÉ DU QUÉBEC À TROIS-RIVIÈRES

IMPACT DU COUVERT NIVAL SUR LES DÉBITS D'ÉTIAGE DANS LES BASSINS NORDIQUES

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RÉSUMÉ

Dans un contexte de prédominance de neige en hiver, plusieurs travaux ont démontré que la durée d'enneigement et la quantité de neige accumulée en hiver dans les bassins-versants des régions froides du monde sont en train de changer en réponse aux changements climatiques, impactant directement les crues printanières dans les cours d'eau (magnitude, fréquence et temps d'occurrence). Plusieurs études ont ainsi montré que les crues printanières seront plus précoces et de moindre amplitude sous un climat plus chaud. Par ailleurs, comparativement peu d'études ont examiné l'impact du cycle nival sur les débits d'étiage (basses eaux) en été. Une partie de la neige qui fond au printemps s'infiltre dans le sol et les aquifères, dont la décharge lente alimente les cours d'eau en été, ce qui permet de maintenir un débit de base, d'assurer un approvisionnement adéquat en eau et de maintenir la qualité des habitats riverains. Il est donc important de comprendre comment les variations dans le stockage de neige impactent le régime hydrologique des rivières en été, afin de mieux anticiper l'impact des changements climatiques sur les ressources en eau.

Dans cette étude, nous chercherons à quantifier comment les variations interannuelles de stockage nival impactent la variabilité de la magnitude et la date d'occurrence du débit d'étiage estival. Une première analyse sera réalisée sur un échantillon de bassins-versants dans le sud du Québec, afin d'offrir un premier diagnostic régional sur cette question et pour développer une méthode simplifiée applicable à un échantillon plus large de bassins et de répondre à la question : est-ce que cette variabilité est conditionnée par le stock de neige maximal accumulé? Dans une deuxième étape, une analyse globale, qui s'inspirera des résultats et méthodes développées dans cette première phase, portera sur un large échantillon de bassins en Amérique du Nord. Des bases de données hydroclimatiques globales seront exploitées à cet effet (ex. Global Runoff Data Centre).

Pour cette étude, des bassins ont été choisis selon plusieurs critères, tout d'abord les bassins à régime hydrologique naturel et situés dans les régions ou le ruissellement est dominé par la fonte de neige avec un minimum de valeurs journalières manquantes. Les variables hydroclimatiques prédictives ayant un effet dominant sur les débits estivaux des bassins versants ont d'abord été identifiées par une analyse de corrélation. La sensibilité des variables de réponse des débits d'étiage en été à chacune des variables prédictives dominantes a ensuite été quantifiée par une analyse de régression linéaire multiple. Les résultats montrent que la pluie accumulée en été contrôle principalement la variabilité interannuelle des débits d'étiages dans les bassins non-montagneux situés plutôt en climat humide. Cependant, l'importance du stock de neige accumulé en hiver contrôle davantage cette variabilité dans les bassins montagneux, ceux plus forestiers et à régime davantage nival. Les hivers avec un équivalent en eau de neige important correspondent à un niveau du débit d'étiage nettement plus important en été. La date d'occurrence des débits d'étiages est plutôt expliquée par la date d'arrivée des pluies automnales dans les régions froides. Ces résultats soulignent l'importance des conditions climatiques hivernales pour les faibles débits en été dans ces bassins versants et offre une autre perspective sur les effets potentiels des changements climatiques sur l'hydrologie des régions froides.

Mots-clés : débits d'étiages, variabilité interannuelle, magnitude, hydrologiques, ÉEN, changements climatiques.

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CHAPITRE I

INTRODUCTION

1.1 Mise en contexte

Les débits d'étiages estivaux saisonniers qui présentent les débits minimaux ou de bases (Roche, 1986), sont importants pour le maintien des écosystèmes et pour répondre aux besoins humains pendant la saison sèche (Godsey et al., 2014). Cependant les différentes projections hydroclimatiques dans les régions froides du monde permettent de prévoir la tendance qui se dessine quant aux impacts du changement climatique sur les évènements hydrologiques extrêmes comme les crues et les étiages. En effet, d'après le Groupe d'experts intergouvernemental sur l'évolution du climat le réchauffement climatique dû aux activités anthropiques est sans ambiguïté un fait très senti de nos jours (GIEC, 2014). Des études menées dans diverses régions du monde indiquent que les changements les plus importants du cycle hydrologique dus au réchauffement sont prévus pour les bassins dominés par la neige (Barnett et al., 2005), car le passage des chutes de neige à la pluie est l'un des effets les plus importants des changements climatiques prévus sur le cycle hydrologique (Zhang et al., 2014). On s'attend ainsi à une réduction des précipitations nivales et à ce que les régimes hydrologiques dits 'nivo-pluvial' se transforment progressivement en régime d'avantage 'pluvio-nival' à pluvial (Berghuijs et al., 2014; Jenicek et al., 2016). Ainsi, avec les effets attendus du réchauffement climatique sur le régime nival, une connaissance précise de l'impact du stockage nival (accumulation de neige en hiver) sur les débits d'étiage en été dans les régions tempérées froides du globe est très importante pour les gestionnaires des ressources en eau. Néanmoins, la relation entre l'hydrologie hivernale et l'hydrologie estivale a été peu étudiée, spécialement au Québec.

1.2 Problématique

Plusieurs travaux ont démontré que la durée d'enneigement et la quantité de neige accumulée en hiver dans les bassins-versants des régions froides du monde est en train de changer en réponse aux changements climatiques (Aygün et al., 2020; Barnett et al., 2005), impactant directement les crues printanières dans les cours d'eau (magnitude, fréquence et temps d'occurrence). En effet, selon Zhang et al. (2014) les précipitations hivernales sous forme de neige alimentent les sources d'eau et contribuent ainsi au débit de pointe printanière. De plus, une étude menée sur des bassins versant aux États-Unis par Berghuijs et al. (2014), a montré qu'une fraction plus élevée des précipitation neigeuses est associée à un débit moyen annuel plus élevé par rapport aux bassins versant avec une fraction de chute de neige plus faible. Ainsi, la diminution de la fraction de précipitation solide sous un climat plus chaud pendant l'hiver peut engendrer une diminution de la magnitude des débits au printemps et affecte la date de fonte des neiges, entrainant ainsi des crues printanières plus précoces et de moindre amplitude (Feng et al., 2007). Par ailleurs, comparativement peu d'études ont examiné la relation entre les débits d'étiage (basses eaux) en été et le cycle nival durant l'hiver. Au Québec per exemple, le débit d'étiage estival a été étudié seulement par rapport aux indices climatiques globaux (Assani et al., 2011; Assani et al., 2010). Une partie de la neige qui fond au printemps s'infiltre dans le sol et les aquifères, dont la décharge lente alimente les cours d'eau en été, ce qui permet de maintenir un débit de base, d'assurer un approvisionnement adéquat en eau et de maintenir la qualité des habitats riverains (Assani et al., 2005; Bradford et al., 2008; Smakhtin, 2001; Walters, 2016). Il est donc important de comprendre comment les variations dans le stockage de neige impactent le régime hydrologique des rivières en été, représenté par les débits d'étiage de 7 jours, afin de mieux anticiper l'impact des changements climatiques sur les ressources en eau. Le choix des débits d'étiage de 7 jours est basé sur ses impact important sur la fonction des écosystèmes aquatiques naturels et sur la dynamique des populations aquatiques et leur impact sur l'activité socio-économique (Assani et al., 2011).

1.3 Objectifs

L'objectif principal de ce projet de recherche est d'examiner l'impact des variations interannuelles de stockage nival (accumulation de neige en hiver) sur les débits d'étiage en été dans les régions tempérées froides de l'hémisphère nord. La question principale guidant cette recherche est : **quelle est la contribution respective de l'accumulation hivernale de neige et de son taux de fonte au printemps à la variabilité interannuelle du débit d'étiage, par rapport à celles de l'évapotranspiration et de la pluie estivale?**

L'hypothèse principale qui sera testée est la suivante :

H0 : Les débits d'étiage ne sont pas ou peu influencés par l'accumulation de neige en hiver mais dépendent plutôt exclusivement des pertes par évapotranspiration et des apports en pluie durant l'été.

H1 : L'accumulation de neige en hiver influe de manière significative sur les débits d'étiage en été.

Ce manuscrit étant un mémoire présenté sous formes d'articles, il s'articule autour de deux objectifs spécifiques bien distinctes :

1) Une première analyse sera réalisée sur un échantillon de bassins-versants dans le sud du Québec, pour quantifier comment les variations interannuelles de stockage nival impactent la variabilité de la magnitude et du temps d'occurrence du débit d'étiage estival afin d'offrir un premier diagnostic régional sur cette question. Le premier article présenté les points méthodologiques détaillés et l'ensemble des résultats issus de ce premier objectif spécifique.

2) Une analyse globale s'inspirera des résultats et méthodes développées dans la première phase et portera sur un large échantillon de bassins dans les régions tempérées-froides de l'hémisphère nord dont les résultats sont présentés dans le deuxième article.

CHAPITRE II

IMPACT OF SNOW STORAGE ON SUMMER LOW FLOW IN SOUTHERN QUEBEC

Article en attente de soumission au journal scientifique *Journal of Hydrology*.

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Abstract

The duration of snow cover and the amount of snow accumulated during winter in the world's cold regions is changing in response to climate change. While reduced snow storage has been frequently associated with earlier and less intense river spring floods, the impact of changing snow storage on summer low flows remains poorly quantified. To address this knowledge gap, the dominant climate predictors of summer low flows were first identified through correlation analysis from 12 tributary basins of the St. Lawrence River in the Canadian province of Quebec. The sensitivity of summer low flow magnitude and timing to each of the dominant predictor variables and basins descriptors was then quantified by hierarchical regression analysis.

The results show that in these low elevation basins, the interannual variability of summer low flows magnitude is sensitive to both winter and summer climate conditions as well as basins characteristics. The correlation results show that summer low flows magnitude is most sensitive to rainfall (rho =0.57), while maximum snow water equivalent (swemax) was the dominant winter control on the magnitude of low flows, particularly at the end of summer (rho = 0.18). As such, winters with a large swemax correspond to a higher level of low flows later in summer. The date of occurrence of low flows is not significantly influenced by snow cover in winter, but rather reflects the timing of autumn rainfall onsets. The hierarchical regression models ($R^2 = 0.78\text{-}0.85$) highlight that interannual variations of summer low flows are not explained only by peak SWE but are induced by a different combination of factors between basins. Rainfall explains most of the interannual variability of summer low flows compared to other climatological variables. The results showed also a positive influence of forest cover on summer low flows ($\alpha = 0.37\text{-}0.42$) compared to agricultural basins. Therefore, further climate warming and snowpack depletion, would have an adverse impact on the ratio of snowfall to total precipitation, so that a reduction in water stored as snowpacks could lead to decreased groundwater recharge and lower summer low flow compared to present conditions.

Keywords: Climate change, low flows, interannual variability, occurrence date, magnitude, snow water equivalent, hydrology.

Introduction

In northern countries such as Canada, the hydrological regime is characterized by long winters dominated by solid precipitation and snow accumulation. Sustained snowmelt leads to a spring freshet, typically the most significant hydrological event of the year and which often results in flooding, especially when combined with rainfall episodes (Buttle et al., 2016; Pomeroy et al., 2016). While a large of part of snowmelt water runs off to streams, another part infiltrates into soils and helps to recharge aquifers, which maintain river baseflow during summer (Maurer et al., 2007; Moore et al., 2007; Stewart et al., 2004) . These seasonal summer base flows, also called low flows, are essential for maintaining ecosystems, water diversion projects, and for ensuring an adequate supply for human needs, especially in cold regions (Burn et al., 2008; Smakhtin, 2001). Thus, it is important to understand the impact of antecedent hydroclimatic conditions and processes that affect the hydrological regime of rivers in summer.

The different hydroclimatic projections in cold regions of the world allow us to predict the emerging trend in the impacts of climate change on extreme hydrological events such as floods and low flows (Diffenbaugh et al., 2013). Studies in various parts of the world indicate that the most significant changes in the hydrological cycle due to global warming are predicted for snow-dominated basins (Aygün et al., 2020a; Barnett et al., 2005). As the shift from snowfall to rain is one of the most important impacts of projected climate change on the hydrological cycle (Zhang et al., 2014), snowpack depletion is expected to occur in several regions and cause a gradual transition from a 'nivo-pluvial' hydrological regime to a 'pluvio-nival' regime in many cold regions (Berghuijs et al., 2014; Jenicek et al., 2016). Whereas increasing precipitation under warming scenarios could lead to enhanced snow accumulation in regions with winter temperatures well below the freezing level, regions with a milder winter climate will experience significant alterations of their hydrological regimes within the next decades (Aygün et al., 2020a; Aygün et al., 2020b).

Given the expected reductions in snowpack volumes under climate change scenarios, a precise knowledge of the impact of winter snow storage on summer low flows is important for water resource managers in cold temperate regions of the world. In general, a low flow is caused by the gradual drawdown of aquifers whose slow discharge feed streams during summer, precipitation deficits in summer, or higher-than-normal evapotranspiration losses, particularly in summer (Smakhtin, 2001). In New England (USA), Hodgkins et al. (2005) showed that about half of the interannual variability in summer low flow is explained by negative anomalies in summer (July-August) precipitation. In China, Tian et al. (2011) showed that the increase in low flow between 1956 and 2004 was related to an increase in annual precipitation averages in the Poyang Lake basin. In Quebec, Canada, Assani et al. (2011) showed that an increase in summer and spring precipitation resulted in an increase in low flow summer over the 1950-2000 period in the southeastern St. Lawrence Basin. Overall, the impact of air temperatures on summer low flow rates appears to be lower than that of summer precipitation, suggesting that an increasing of evapotranspiration losses due to an increasing of air temperature would lead to a reducing (negative correlation) summer low flow volume (Hodgkins et al., 2005; Yang et al., 2002).

Several studies have shown that the duration of snow cover and snowpack volume in cold regions watersheds is changing in response to climate change, directly influencing the magnitude, timing and frequency of spring floods (Curry et al., 2018; Zhang et al., 2014). These studies have analysed the relationship between the characteristics of spring high flows in cold regions and their relationship with snow cover and climate indices. Zhang et al. (2014) showed in their study of alpine watersheds in Asia (1960-2007) that winter and summer accumulated temperature and the amount of snowfall were significantly correlated with the spring snowmelt peak discharge, whereas other climatic factors had no direct effect on the spring freshet. Also, extensive work in the western United States (Berghuijs et al., 2014; Stewart et al., 2005) has clearly shown that a higher fraction of precipitation falling as snow is associated with a higher mean annual discharge and that warmer air temperatures in spring result in a higher fraction of the annual discharge occurring earlier in the hydrological year. Many studies throughout the world have reported that spring floods in snow-affected basins will be earlier and of lesser magnitude in a warmer climate (Bavay et al., 2009; Blöschl et al., 2017; Boyer et al., 2010; Buttle et al., 2016; Guay et al., 2015; Veijalainen et al., 2010). However, while the impacts of snowpack volume on spring discharge has been widely documented, the relationship between winter and summer in cold regions has been comparatively little studied.

Only a few studies have investigated the link between snowpack characteristics and interannual variability of low flows, finding in general that winter snow accumulation can affect low flows in the following summer. In the Swiss Alps snow water equivalent was found to be the dominant winter predictor of 7-day low-flow (Jenicek et al., 2016). Similar findings have been reported from Sierra Nevada basins in the western United States, where Godsey et al. (2014) found that the decrease in maximum snow water equivalent leads to a decrease in low flows during summer. Similar results were found in the maritime western U.S. mountains (Cooper et al., 2018) and in the mountains of western North American (Dierauer et al., 2018), where dry winters lead to significantly lower low flows in summer. Other studies showed that the variability of summer low flows could be explained by not only by seasonal Melt in spring but also by glacier wastage in the basin. For exemple, the presence of glaciers (Schaefli et al., 2005; Comeau et al., 2009) can impact low flows due to the contribution of glacier melting during summer. Almost all studies have been conducted in mountain catchments in western North America or in the Swiss Alps, non-mountainous basins have not been studied. The interannual variability of summer low flows has been studied in Canada in relation to atmospheric and oceanic circulation climate indices (Assani et al., 2011; Assani et al., 2010; Bonsal et al., 2008), but the hydroclimatic variables responsible for the interannual variability in the magnitude and timing of summer low flows have not yet been identified. The present study intends to fill this gap by estimating the relative impacts of interannual variations in snow storage, summer rainfall and evapotranspiration on the variability of summer low flows in southern Quebec, Canada. We further investigate the relationship between low flow and basins biophysical characteristics to explore regions that might become more sensitive to climate change in the future.

Data and methods

Study sites

12 tributary basins of the St. Lawrence River in southern Quebec were selected for the study (Fig. 1). Basin areas range from 366 to 4,504 km^2 with elevations less than 850 m (Table 1). The choice of basins was based on the length of the flow measurement record, a hydrological regime as close as possible to natural conditions (i.e., near-natural flows with no major human influences, such as dams or water diversions), and the presence of snow monitoring stations within or close to the basin. These snow monitoring stations were used in a former study to calibrate and validate the GR4J-Cémaneige hydrological models in the 12 basins (Nemri et al., 2020). The selected basins are spatially distributed between the north and south shores of the St. Lawrence River, within four homogeneous hydrological regions used by the Quebec Center of Water Expertise (CEHQ), the governmental organization in charge of hydrological monitoring and forecasting in Quebec. Basins in the northwestern St. Lawrence region (Matawin, Batiscan, and Bras du Nord) straddle the Canadian Shield with a more rugged and forested landscape, and the mostly agricultural St. Lawrence lowlands, and is characterized by a continental climate. The northeastern St. Lawrence region (Godbout basin) is on the Canadian Shield and has a maritime climate, while basins in the southwestern St. Lawrence region (Nicolet, Acadie) straddle the Appalachian Mountains with a forested landscape, the St. Lawrence lowlands and have a maritime climate. Basins in the southeastern St. Lawrence region (York, Beaurivage, Bécancour, Famine, Etchemin, Ouelle) are in the Appalachian region and characterized by a mixture of maritime and continental climate (Assani et al., 2010; Mazouz et al., 2013). Among the studied catchments, only Acadie is dominantly agricultural (73%), while the other basins are dominantly forested, i.e., more than 70% of their area is covered by forest (Table 1).

Hydrometeorological data

This study uses continuous meteorological data extracted from daily climate grids developed by the Atmospheric Environment Information Service (SIMAT) in collaboration with the Quebec Center of Water Expertise (CEHQ) (Bergeron, 2015). Basin averages of mean daily air temperature and total precipitation were calculated from the grids (Nemri et al., 2020). Daily streamflow records at the outlet of the 12 basins were obtained from the CEHQ (www.cehq.qc.ca). More than 30 years of records were chosen for analysis to match the standard record length used for defining climate normals (Table 1). This period is limited by the gridded climate product extending from 1961 to 2015. Potential evapotranspiration (ETP) was calculated according to Oudin (2014) from daily air temperature and elevation. We use mean (basin-wide) snow water equivalent (SWE) for the 12 study basins previously simulated by Nemri et al. (2020) with the Cémaneige snow model (Valéry et al. 2014). Rain-snow partition, snowpack accumulation and snowmelt were calculated at five equal-area elevation bands, and then averaged to the whole basin. ETR and river discharge were simulated with the GR4J model (Perrin et al., 2003) with ETP, snowmelt and rainfall as inputs. Both discharge and SWE observations were used in a multi-objective calibration of the GR4J-Cémaneige model by Nemri and Kinnard (2020). SWE observations come from bi-weekly surveys made by the Ministry of Environment and Fight against Climate Change (MELCC) in each of the studied basin and represent the average of 10 manual measurements made with a snow corer at fixed points along a 3 × 100-meter trail (MELCC, 2008). Model results yielded a good performance on independent observations, with Nash-Sutcliffe efficiency (NSE) coefficients ranging from 0.67 to 0.90 (median: 0.80) for discharge and 0.20 – 0.78 (median: 0.70) for SWE. Using the modeled SWE instead of SWE observations circumvents the problem of the sparse temporal (bi-weekly) and spatial (one or few points) coverage of the observations. Further model description and details on forcing data, calibration and validation are given in Nemri et al. (2020).

Antecedent hydroclimate conditions and statistical analysis of summer low flow

The minimum daily flow measured over seven consecutive days between May and September was used as summer low flow variable. 7-day running streamflow averages were used to minimize the effect of potential outliers. This 7-day low flow indicator (Q7) is widely used as it is not susceptible to temporary upstream flow changes that can affect one-day low flow calculations (Riggs, 1985).

Seven variables related to winter, spring and summer antecedent meteorological and water storage conditions in the basins were selected as potential predictors of the minimum summer low flow (Q7min) (Table 2). The advantage of this choice of predictors is that only daily data of snow water equivalent (SWE), snow cover fraction in the basin, precipitation, air temperature, and streamflow are required for their calculation. Winter conditions were represented by the maximum (peak) accumulated SWE in the basin (swemax) and its date of occurrence (swemaxdoy). Spring conditions are represented by the date of snow disappearance (meltoutdoy), calculated as the first day following the peak SWE date with near snow-free conditions, i.e. with a snow cover less than 10% in the basin (Jenicek et al. 2016). The snow cover fraction was simulated by the Cémaneige model (Nemri et al., 2020; Valéry et al., 2014). It is hypothesized that a thicker and later snowpack, as well as a longer snow cover duration could favour aquifer recharge and help sustaining higher low flows later in summer. In addition, the average melt rate (meltrate) was calculated for the period between the date of maximum SWE (t0) and the meltoutdoy (t1), following equation (1) (Barnhart et al., 2016):

$$meltrate = \frac{\sum_{t0}^{t1} \Delta SWE_t}{D} \tag{1}$$

where ΔSWE_t represent the sum of changes in SWE (ablation) and D the number of days with negative ΔSWE_tbetween peak SWE date (t0) and complete snow disappearance (t1).

Three summer predictors were considered. In the St. Lawrence valley, the low flow regime is typically interrupted by the onset of sustained rain events in the fall, which varies between years depending on synoptic patterns, in particular the southern migration of the polar front in the fall which favours cyclogenesis and frontal precipitation events over southern Quebec. As such, the date of minimum summer flow (Q7mindoy) was used as predictor of its magnitude (Q7min) to represent this phenomenon. The accumulated actual evapotranspiration (ETR) and rainfall (RAIN) between the date of snow disappearance (meltoutdoy) and the date of minimum flow (Q7mindoy) were also calculated in each basin. Taken together these two variables determine the amount of available water in a given year and basin after snowmelt.

We first used Spearman's rank correlation to explore bivariate relationships between each potential predictor and summer low flow. The nonparametric Spearman correlation was used because several variables are not normally distributed. The analysis was first conducted for all basins combined (global analysis) and then separately for each basin. For this purpose, predictor and response variables were previously standardized to zero mean and unit standard deviation per basin, to allow pooling basins on a common unit scale for the global analysis. To investigate memory effects of snowpack on low flow, correlations were calculated between the maximum SWE and the minimum flow within a 10-day running window from June to September in each basin.

Multilevel or 'hierarchical' regression was then used to describe the combined effect of antecedent hydrometeorological variables on summer low flows interannual variability, while also considering the effect of basins properties. Several basin biophysical descriptors were also considered, i.e. forest cover, drainage density, mean basin slope and basin area (Table 1). Hierarchical regression is a widely used statistical model used when observations are organized (grouped) at more than one level and which allows to jointly model the within- and between-group variability (Gelman and Hill 2006). The model relates the response variable (Q7min) to hydroclimatic predictors (first level fixed effects) with basin (ID) as grouping variable (random effect) and basin descriptors as group-level variables (second level fixed effects). The hierarchical model is represented by the following equations (Gelman and Hill 2006, p.266):

$$y_t = N\left(\alpha_{j[t]} + \beta x_t, \sigma_t^2\right), for\ t = 1, \ldots, n \tag{2}$$

$$\alpha_j = N\left(\gamma_0 + \gamma_1 u_j, \sigma_\alpha^2\right), for\ j = 1, \ldots, J \tag{3}$$

where x_t and u_t represent predictors at the annual levels, n represents the number of annual observations in each basin j for a total of J = 12 basins. β is the vector of regression slopes for hydroclimatic predictors (first-level fixed effects), $\alpha_{j[t]}$ is the intercept for basin j, γ_0 is the global or mean intercept value, γ_1 is the vector of regression slopes for basin descriptors (second level fixed effects), and σ_t^2 and σ_α are the standard deviations (model errors) at the first and second levels, respectively. *N* symbolizes the normal distribution.

The data were first standardized (Z score) per basin in order to common scale in the analysis and transformed to a normal distribution using BoxCox transformation (Sakia, 1992) when necessary. Predictor multicollinearity was tested by calculating the variance inflation factor (VIF), discarding predictor variables with a VIF larger than 10 (O'brien 2007). Separate hierarchical regression models were fitted for each individual summer month, from June to September, as well as for the whole period. Hence for individual months Q7min is the minimum 7-day discharge over the month, while for the whole period Q7min is the minimum 7-day discharge over the whole summer period (June-September). In all cases RAIN and ETR are summed between meltoutdoy and Q7mindoy. Three different models of increasing complexity were considered: (1) only considering hydroclimate predictors, without basin grouping effect; (2) hydroclimate predictors and a basin grouping effect; (3) hydroclimate predictors with basin grouping and basin characteristics as basin-level predictors. The best model was selected based on the Akaike Information Criterion (AIC) (Akaike, 1987). All data processing and statistical analysis was carried out with the MATLAB and R software.

Results

Correlation between selected predictors and response variables

The global correlation heat map shows the strength of the association between all predictor and dependent variables when pooling all basins together (Fig. 2). All variables related to snow conditions are significantly ($p < 0.05$) and positively correlated. The snow disappearance date (meltoutdoy) is positively and significantly correlated with peak SWE (swemax, rho = 0.49) and its date of occurrence (swemaxdoy, rho = 0.47). In addition, the melt rate is strongly correlated with swemax (rho = 0.63) and its date of occurrence (swemaxdoy, rho = 0.79). Hence, thicker snowpacks tends to occur later and also to last longer in the spring. This also explains the higher average melting rate occurring when snow persists under warmer spring conditions, as found elsewhere (Musselman et al. 2017). Correlation between winter conditions and summer variables were generally low and often nonsignificant, as expected. However, significant and inverse correlations are observed, notably between meltoutdoy and ETR and RAIN, which occurs because meltoutdoy controls the period over which the sum of rainfall and ETR is calculated.

Fig. 2 allows a first evaluation of the global relationship between summer low flow (Q7min) and antecedent hydroclimate conditions. A weak, but significant positive correlation (rho = 0.18, $p < 0.05$) is found between peak SWE and Q7min for the 12 pooled basins. Additionally, both meltoutdoy (rho = 0.09, $p < 0.05$) and meltrate (rho = 0.11, $p < 0.05$) are weakly but positively correlated to summer low flow. Thus, meltoutdoy and melting rate do not explain much of the variability in low flow magnitude in summer. The positive correlation between meltrate and Q7min is contrary to the hypothesized increased recharge under slower melt rates and could occur due to the collinearity between swemax and meltrate. As expected, summer preconditioning of low flow conditions is more important than winter and spring preconditioning: the amount of rain accumulated between the snow disappearance date and the summer low flow date (RAIN) is the best overall predictor of Q7min (rho = 0.27), so that wetter summers globally lead to higher summer low flows. Evapotranspiration losses (ETR) are associated with reduced low flows, but the correlation is weaker (rho = -0.11) than that for RAIN and swemax.

No significant correlations appear between winter (snow) variables and the timing of summer low flow (Q7mindoy). In contrast, Q7mindoy is strongly and positively correlated with RAIN (rho = 0.82) and ETR (rho = 0.83). However, these correlations arise because later low flows increase the period over which RAIN and ETR are calculated. Also, no global correlation emerges between Q7min and its date of occurrence (Q7mindoy).

Impact of peak SWE on low flow during summer

To explore in more details the memory effect of snowpack conditions and how long water from snowmelt contributes and affects low flow in summer. on summer low flows, the peak SWE (swemax) was correlated with the minimum 7-day low flow every 10 days for the 12 selected basins separately (Fig. 3), similar to the approach used by Jenicek et al. (2016) for catchments in Switzerland. The 10-day period was chosen to be large enough to capture flow recessions while keeping a high temporal resolution for the analysis. These 10 days correlation coefficients show the temporal sensitivity of low flow conditions to preceding winter accumulated snowpack. The influence of peak SWE on summer low flow tends to increase over time and peaks between late August and mid-September, with however heterogeneity between basins (Fig. 3). Two basins stand out from this general trend: the northernmost basins Godbout (ID1) and York (ID2), for which SWE is rather weakly but negatively correlated with late summer Q7min. The peak SWE is also negatively correlated with Q7min early in summer in five basins (Etchemin, Bécancour, Ouelle, Famine, Nicolet) such that larger peak SWE is associated with a lower Q7min in early/mid-June. The snow disappearance date (meltoutdoy) typically occurs in April-May, varying by ±1-2 weeks each year in a given basin and increasing towards higher latitudes. In contrast, the occurrence date of Q7min occurs on average from mid August to mid September, without any latitudinal trend, but with large interannual variability (~ ±1 month).

When compared across all 12 catchments, the correlation between the preconditioning variables and Q7min in September, when the correlation with peak SWE is highest (Fig. 3), shows that the minimum discharge is positively correlated with peak SWE in 10 of the 12 basins, with correlations between 0.21 and 0.41 (Fig. 4). However only five basins pass the significance threshold ($p < 0.05$). The smaller number of observations in individual basins explains the higher threshold correlation for statistical significance, compared to the pooled basins analysis in Fig. 2. Despite the significance of the correlations, their values are generally not strong, suggesting that swemax plays a secondary role on low flow variability compared to summer conditions. The correlation between swemax and September Q7min also displays significant inter-basin heterogeneity. Other winter predictors (swemaxdoy, meltoutdoy and meltrate), are most often positively correlated with Q7min, but are rarely significant, except in Acadie and Batiscan.

Fig. 4 highlights a strong and significant correlation between Q7min and accumulated rainfall (RAIN) in late summer (September) in all the studied basins. The correlation coefficients vary between 0.28 in Famine and 0.62 Godbout. Conversely, the accumulated ETR has significant negative effects on late summer Q7min only in three basins (Godbout, Bras du Nord and Batiscan) while a positive effect is found in three other basins (Ouelle, Etchemin and Beaurivage). While higher ETR should in theory lead to reduced low flow, the positive correlation between RAIN and ETR (Fig. 2) complicates the interpretation of simple bivariate relations. To further examine the combined influence of rainfall and ETR on low flow, Q7min was correlated with the net rainfall (NETRAIN), i.e. the available water for runoff. The correlation between Q7min and net rainfall either increases or decreases slightly compared to the correlation with rainfall alone, suggesting that the magnitude of Q7min is more related to rainfall than evapotranspiration in summer. Interestingly, Q7min is positively and significantly correlated with its date of occurrence (Q7mindoy) in four basins, suggesting that later low flow in September have a higher magnitude, contrary to expected when river flow is dominantly fed by aquifer discharge. These results highlight that to better discriminate the influence of the respective hydroclimate predictors, their multivariate influence on low flow must be assessed, which is explored in the next section.

Low flow response to multivariate antecedent hydroclimate conditions

As highlighted in the previous section, Q7min in summer can be related to a combination of preconditioning factors, with heterogeneity between basins. Thus, a hierarchical regression model was developed to investigate the combined effects of hydroclimate predictors, take into account the potential influence of catchment characteristics on summer minimum discharge, and to better tease out the memory effect of snow cover on summer discharge. Collinearity tests using the variance inflation factor (VIF) showed that the meltoutdoy had a VIF larger than the rule of thumb value of 10 for collinear variables

(O'brien, 2007) and was thus remove for the predictor set. As shown in Fig. 2, the snow disappearance date is correlated with the maximum SWE and the melt rate, as a thicker snowpack tends to last longer into the spring and increases the mean melt rate.

The varying-intercept model (model 2), which accounts for random variations in Q7min between basins, significantly improves the performance of the model compared to using fixed hydroclimatic effects only (model 1) (Table 3). This was expected, given the basin heterogeneity in Q7min correlations highlighted in Fig. 2 and 3. Adding basin descriptors in model 3 further improve the predictive performance relative to model 2, suggesting that part of the inter-basin random variations can be explained by differences in basin characteristics (Table 3). This model was thus retained as the best model to explain Q7min interannual variations (AIC = 611.77). Adding interactions between predictors (fixed effects) did not improve the model further.

Results for the hierarchical model (model 3) are presented in Table 4. Fixed effects, i.e. first-level hydroclimate variables and second-level basin descriptors, explain 47-54% of the interannual variability in Q7min (marginal R^2 = 0.47-0.54) while random (inter-basin differences) effects account for 28-38 % of the variance in Q7min (conditional R^2 = 0.78-0.85). It can be observed that the relation between swemax and Q7min for the whole summer is similar to the relation seen in September, which is because Q7min occurs on average in late summer (Table 4). Swemax has a positive influence on Q7min throughout summer (α = 0.15-0.25, $p < 0.05$), in accordance with the bivariate correlation analysis. The influence is highest in June and August. Swemaxdoy is also positively related to Q7min in June-September (α = 0.14-0.19, $p < 0.05$), so years with delayed melting and a later peak SWE tend to result in higher minimum discharge at the end of summer. The melt rate has a negative influence on Q7min in June (α = -0.15) and August (α = -0.18). This is interpreted to reflect the fact that slower melting of the snowpack favours infiltration, which in turn would support subsurface drainage contributions to baseflow in early summer, and aquifer recharge which would favor a delayed contribution to low flows in August.

The accumulated rainfall following snow cover disappearance (RAIN) has a strong and sustained influence on Q7min throughout summer (α = 0.44-0.59). Evapotranspiration losses (ETR) have a negative influence on Q7min (α = -0.18 to -0.39), decreasing towards the end of summer. Thus, the expected negative effect of ETR on Q7min stands out clearly in the multivariate analysis compared to bivariate correlations (Fig. 2-3), since the simultaneous effect of the other predictors is taken into account in the hierarchical models. It is worth noting that while ETR has a lesser influence than rainfall on low flow, the influence of SWE on Q7min is of similar magnitude than that of ETR in late summer (August-September). In addition, and contrary to bivariate correlation results (Fig. 2-3), the multivariate analysis showed that the interannual variations in low flows timing (Q7mindoy) has a negative significant influence on Q7min volume overall in summer except in September. This is consistent with the notion that later flows have lower magnitude due to the progressive drawdown of aquifers feeding baseflow.

Inter-basin heterogeneity in the hydroclimate preconditioning of low flows was already noted in previous section and Fig. 3 and 4. The hierarchical model analysis yields interesting insights on the magnitude and causes of this heterogeneity. Partial residual plots of the significant winter/spring predictors are shown for the summer period in Fig. 5 and for the month of August in Fig. 6. The varying intercept coefficients, visible on these figures, indicate significant differences in mean low flow conditions amongst basins that are not explained by hydroclimate factors. Among the basin-level descriptors included to explain this inter-basin variability, the drainage density and basin slope were found to have no influence on Q7min, while the forest cover has a significant positive influence on Q7min from July to September (α = 0.37 - 0.38) and more over the whole summer period (α = 0.42) (Fig. 7). Hence the percentage of forest cover in a basin appear as a significant condition that leads to higher minimum flow, especially towards late summer. Still, significant random inter-basin variability in Q7min remains, that is unexplained by hydroclimate predictor and basin descriptors, as shown by the random effect variances (τ_{00} = 0.32 - 0.34) and corresponding conditional R^2 in Table 4.

Discussion and conclusion

Our results highlighted the impact of antecedent hydrological conditions on summer low flow in twelve southern Quebec rivers with a natural regime and significant variability in winter and summer precipitations. The global bivariate correlation analysis (cf. section 3.2) showed that snow conditions in winter can affect low flow the following summer, with maximum SWE showing the clearest positive association with Q7min. However, the basin-level correlation analysis showed significant heterogeneities among these basins. Overall, the bivariate correlation results show that the accumulated winter snowpack affects groundwater recharge and thus river flow during dry summer periods, as found in other snow dominated areas (eg. Beaulieu 2012). This is in accordance with the results found in mountainous basins: in the Sierra Nevada in western USA (Godsey et al., 2014), in the Swiss Alps (Jenicek et al. 2016), and in Austria (Laghari et al., 2012) where SWE showed the best predictive ability for the 7-day low-flow among all winter-related predictors. Our work demonstrates that across most of studied basins, the effect of maximum SWE on Q7min is more important towards the end of summer (August/September). This delayed response to snowmelt is interpreted to result from the predominant recharge of aquifers during spring snowmelt followed by the progressive outflow of aquifers to streams in summer, as documented in several cold regions (Aygün et al., 2020b; Boumaiza et al., 2020; Jasechko et al., 2014). This is different than the situation in low elevation Alpine and pre-Alpine catchments, where the sensitivity of Q7min to SWE was found to decrease over the course of summer (Jenicek et al., 2016). This could explain that the transfer modes of "snow anomalies" are different, i.e. via the slow discharge of aquifers in Quebec, and via faster subsurface runoff in alpine basins. However, for two basins (York, Godbout), a thicker snowpack in winter was related to lower low flows volume in summer. This is possibly because the basins have a maritime climate, with year-round precipitation abundance, which could explain the low dependence of streamflow on aquifer recharge by snowmelt.

The hierarchical regression approach used in this study allowed examining the combined effect of hydroclimate predictors while accounting for inter-basin variability in low flow. Our results show that both summer climate (rain and evapotranspiration) and winter conditions (snow storage and melt dynamics) influenced summer low flows interannual variations. Cumulative summer precipitation is the dominant control on summer low-flow variability, with evapotranspiration losses and snow storage play significant but secondary roles. Similar results regarding a large influence of rainfall regimes on low flows have been documented for historical river summer low flows in New England (Hodgkins et al., 2005), for lower elevation basins in Switzerland (Jenicek et al., 2016) and southeast St. Lawrence watershed in Quebec (Assani et al., 2011). While interannual variations in rainfall is unsurprisingly the first controls on summer low flows, antecedent winter conditions still represent an important source of interannual variability, of similar magnitude than ETR losses.

Differences are found between bivariate correlation results (global and by basin) and the multivariate analysis. While some predictors (swemax and RAIN) have consistent relationships with Q7min, partly because these two variables are not related, other variables have different effects when the influence of the other predictors is taken into account, which is not the case in the bivariate analysis. As such, ETR and Q7mindoy have more consistent negative relationships with Q7min once the other predictors are taken into account in the regression models.

Our results showed significant inter-basin variability in low flow which was partly explained by land cover (forest cover) (Table 4), whereas topography and drainage density which both favour faster water transfers within basins were unimportant, contrary to results found by Jenicek et al. (2016) in alpine basins. The influence of forest cover can be explained by the more porous forest soils, which favour the infiltration of snowmelt water, subsurface flow processes and aquifer recharge compared to agricultural lands with frequently compacted and clay-rich soils in Quebec. As a result, the baseflows in forest-dominated basins are prolonged, so that Q7min occur later in summer with higher volumes (Neary et al., 2009). This indicates the importance of groundwater contribution to summer low flows in the region. These results are in agreement with those documented in British Columbia forested basins (Beaulieu et al., 2012) that showed the important contribute of recharged subsurface reservoir by spring snowmelt to support a more stable summer base flow. However, many studies in North America (Caissie et al., 2002; Pike et al., 2003), some of them conducted in Quebec (Mumma et al., 2001; Quilbé et al., 2008) have demonstrated that deforestation in a basin, through forest harvesting, leads to an increase in low flows in general. The differences between our results and previous work in other cold region basins may be partially explained by the differences in the studied catchments properties for basin and in the hydrological regime. Thus, a short-term basins deforestation can keep its porous soils but decreasing the losses per ETR thus the magnitude of low flows in these basins are higher than in agricultural one which characterised by a compacted soil.

Our results showed a significant sensitivity of summer low flow to antecedent snow storage. Although we did not explore the implication of climate change on snow cover, rising temperatures are expected to decrease snowpacks and snow cover (Sturm et al., 2017). In mountains this effect will depend on elevation (rise of the isotherm zero) (Jenicek et al., 2018), but in non-mountainous landscapes such as in the St. Lawrence valley climate warming will cause a progressive widespread shift from snowfall to rainfall (Aygun et al., 2020b). A decrease in snow/precipitation ratio would lead to an increase in the volume of water stored as the snowpack (Boyer et al., 2010). Based on our findings, the reduced snow storage anticipated in response to climate change could also reduce the efficiency of groundwater recharge, which would reduce low flow conditions in late summer and negatively impact ecosystems integrity and ecological services.

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Table 1. Characteristics of the 12 selected study basins. Basins are ranked by latitude, from north (ID=1) to south (ID=12).

Basin ID	River	Area (km^2)	Data from (to 2015)	Mean slope (%)	Drainage density (km^{-1})	Forest cover (%)	Elevation range (m)
1	Godbout	1577	1974	3.87	0.486	99.5	300-400
2	York	647	1980	4.81	0.532	100	400-500
3	Ouelle	769	1982	1.51	0.387	97.4	300-400
4	BrasduNord	646	1965	4.67	0.602	100	500-600
5	Etchemin	1152	1980	2.46	0.54	74.5	300-400
6	Matawin	1387	1931	2.5	0.456	96.9	400-500
7	Beaurivage	708	1925	1.17	0.62	61.7	100-200
8	Batiscan	4504	1967	3.01	0.434	92.6	300-400
9	Bécancour	2163	1999	2.31	0.547	74.1	200-300
10	Famine	696	1964	1.5	0.484	87.4	300-400
11	Nicolet	1550	1966	1.89	0.516	40	200-300
12	Acadie	367	1979	0.21	0.715	25.7	100-200

Table 2. Predictors and response variables used in statistics analysis.

Variables	Abbreviation (Units)
Response variable	
Minimum of 7-day moving average of discharge	Q7min (mm)
Winter predictor variables	
Maximum SWE in winter	swemax (mm)
Occurrence date of maximum SWE in winter	swemaxdoy (day of year)
Snow disappearance date	meltoutdoy (day of year)
Rate of melting	meltrate (mm/day)
Summer predictor variables	
Accumulated summer precipitation	RAIN (mm)
Occurrence date of 7-day minimum discharge	Q7mindoy (day of year)
Accumulated actual evapotranspiration	ETR (mm)

Table 3. Hierarchical regression models tested.

Models	AIC
Model1: swemax + ETR + RAIN + swemaxdoy + meltrate + Q7mindoy	1183.15
Model2: swemax + ETR + RAIN + swemaxdoy + meltrate + Q7mindoy + (1 | basin)	614.56
Model3: swemax + ETR+ RAIN + swemaxdoy + meltrate + Q7mindoy + mean_slope + drainage_density + forest_cover + area + (1 | basin)	611.77

Table 4. Coefficients and variance components for monthly and whole summer hierarchical regression models of summer low flow (Q7min). '-' indicates non-significant terms ($p > 0.05$). σ^2: residual variance (model error); τ_{00}: random effect (inter-basin) variance in Q7min. Marginal R^2: variance explained by fixed effect; conditional R^2: variance explained by random and fixed effects.

Predictors	Q7min (June)	Q7min (July)	Q7min (Aug.)	Q7min (Sept.)	Q7min (summer)
Intercept	-	-	-	-	-
Winter conditions					
swemax	0.24	0.19	0.25	0.15	0.15
swemaxdoy	0.19	0.15	0.16	0.14	-
meltrate	-0.15	-	-0.18	-	-
Summer conditions					
ETR	-0.39	-0.34	-0.19	-0.18	-0.29
RAIN	0.44	0.57	0.59	0.51	0.57
Q7mindoy	-0.19	-0.19	-0.2	-	-0.27
Descriptors					
area	-	-	-	-	-
drainage_density	-	-	-	-	-
forest_cover	-	0.37	0.38	0.37	0.42
mean_slope	-	-	-	-	-
σ^2	0.2	0.21	0.21	0.25	0.18
τ_{00}	0.35	0.33	0.32	0.33	0.44
Marginal R^2 / Conditional R^2	0.49 / 0.78	0.52 / 0.82	0.54 / 0.82	0.53 / 0.82	0.47 / 0.85

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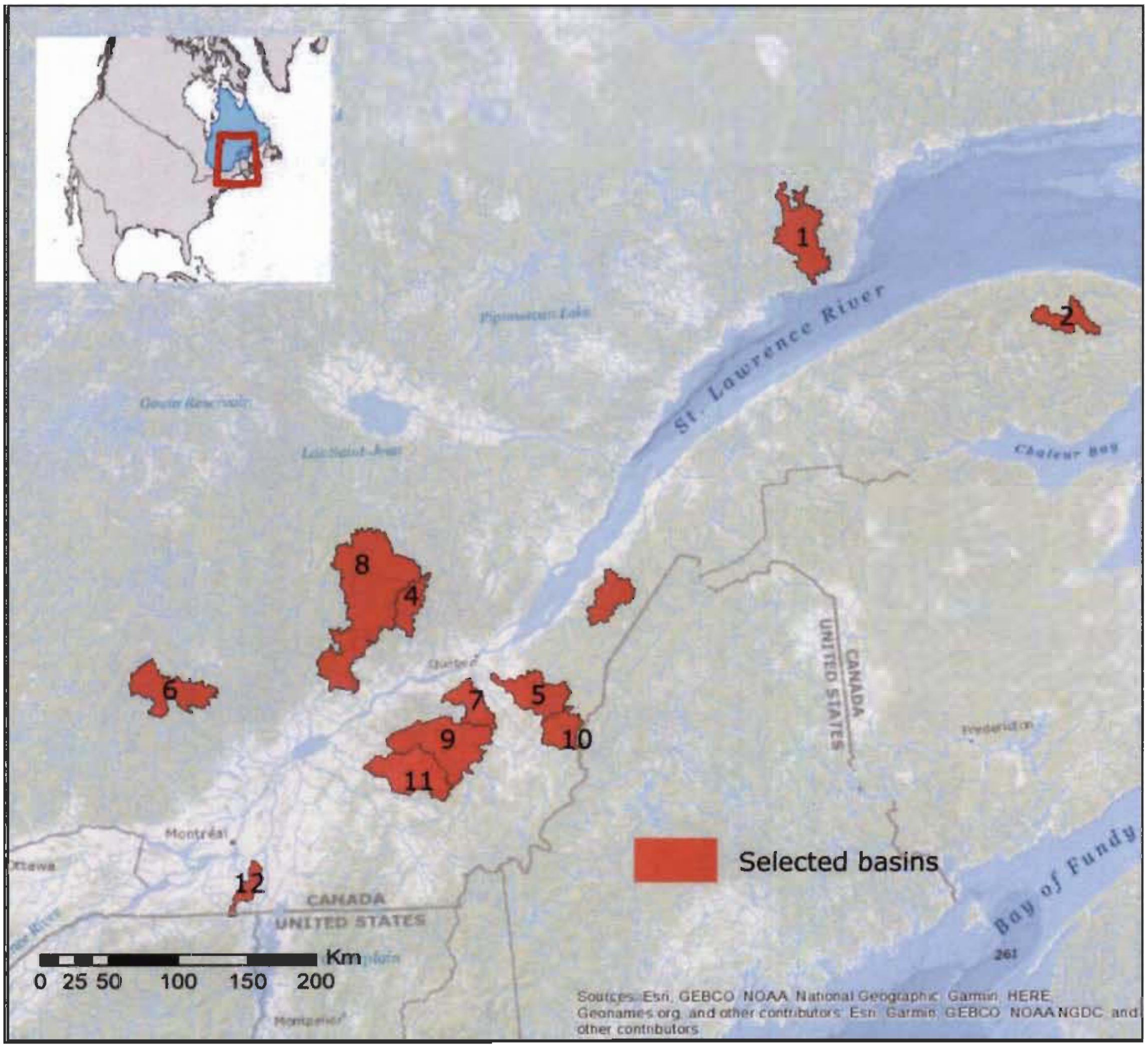

Fig. 1 The 12 tributaries basins of the St Lawrence River selected for this study. Basins are delineated in red. 1 Godbout, 2 York, 3 Ouelle, 4 Bras du Nord, 5 Etchemin, 6 Matawin, 7 Beaurivage, 8 Batiscan, 9 Bécancour, 10 Famine, 11 Nicolet, 12 Acadie.

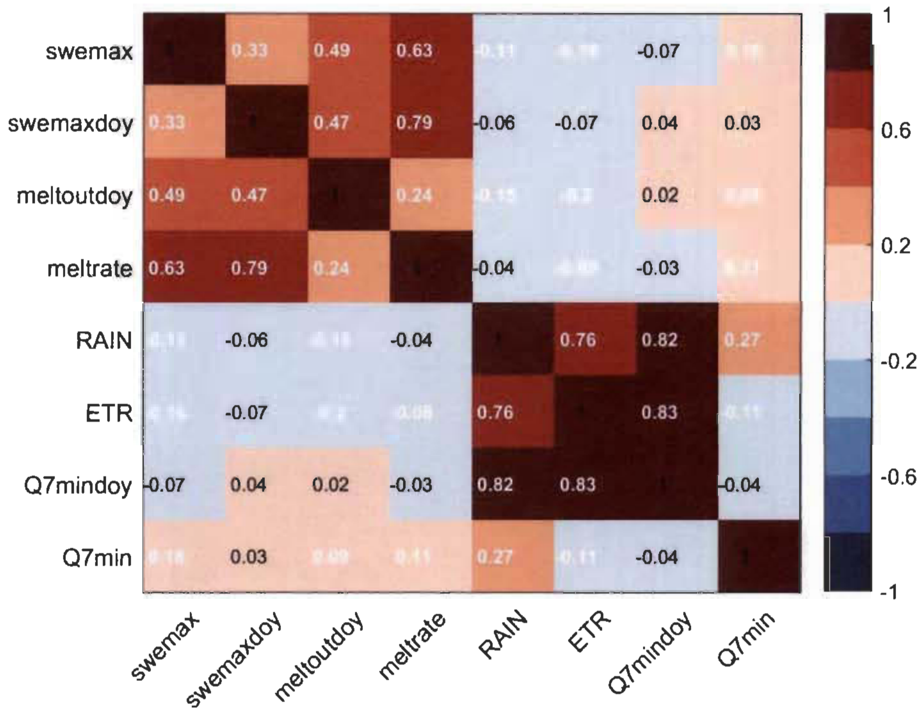

Fig. 2 Heat map showing Spearman rank correlation coefficients (rho) between all predictor and response variables for the 12 basins pooled together. Significant correlation ($p < 0.05$) are indicated in white. The red (blue) colour scale indicates positive (negative) correlations.

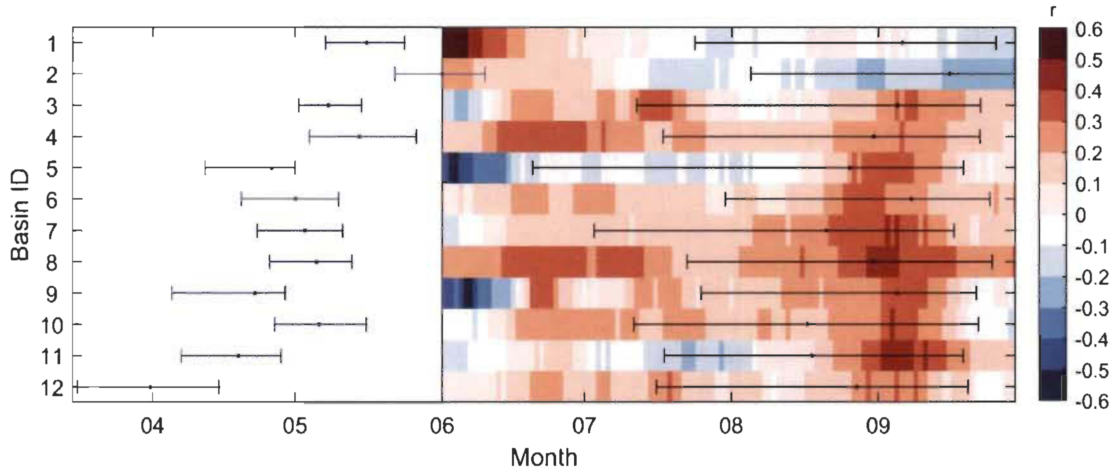

Fig. 3 Dependence of 7-day minimum discharge on maximum SWE for all studied catchments :1 Godbout, 2 York, 3 Ouelle, 4 Bras du Nord, 5 Etchemin, 6 Matawin, 7 Beaurivage, 8 Batiscan, 9 Bécancour, 10 Famine, 11 Nicolet, 12 Acadie. Basins IDs are ranked from north to south. Blue bars represent the variability in snow disappearance date (meltoutdoy) and black bars that of summer low flow occurrence date (Q7mindoy). Whiskers represent 10 and 90% percentile. Red colours indicate a positive effect of SWE on minimum discharge (positive correlation) while blue colours indicate a negative effect (negative correlation).

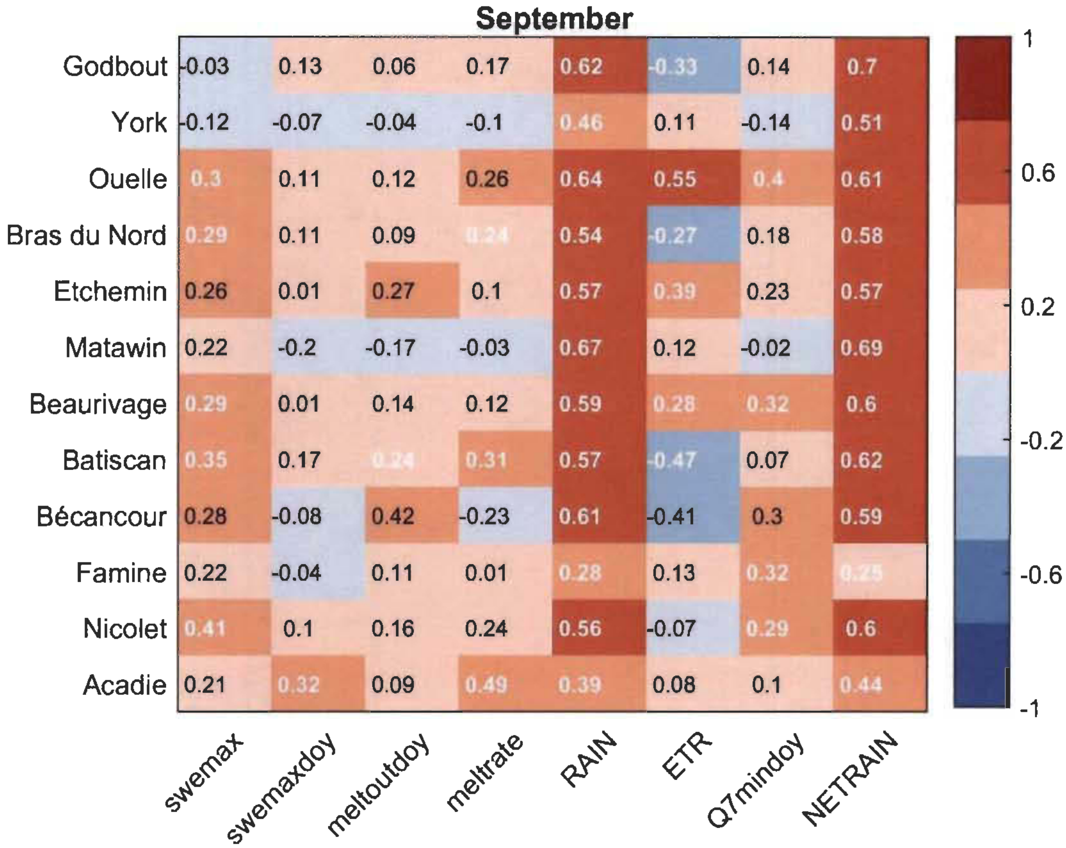

Fig. 4 Heat map showing Spearman correlation coefficients between hydroclimate predictors (columns) and minimum summer low flow magnitude (Q7min) for the twelve basins (rows) in September. Significant correlation ($p < 0.05$) are highlighted in white; blue colors indicate negative correlations and red colors positive correlations.

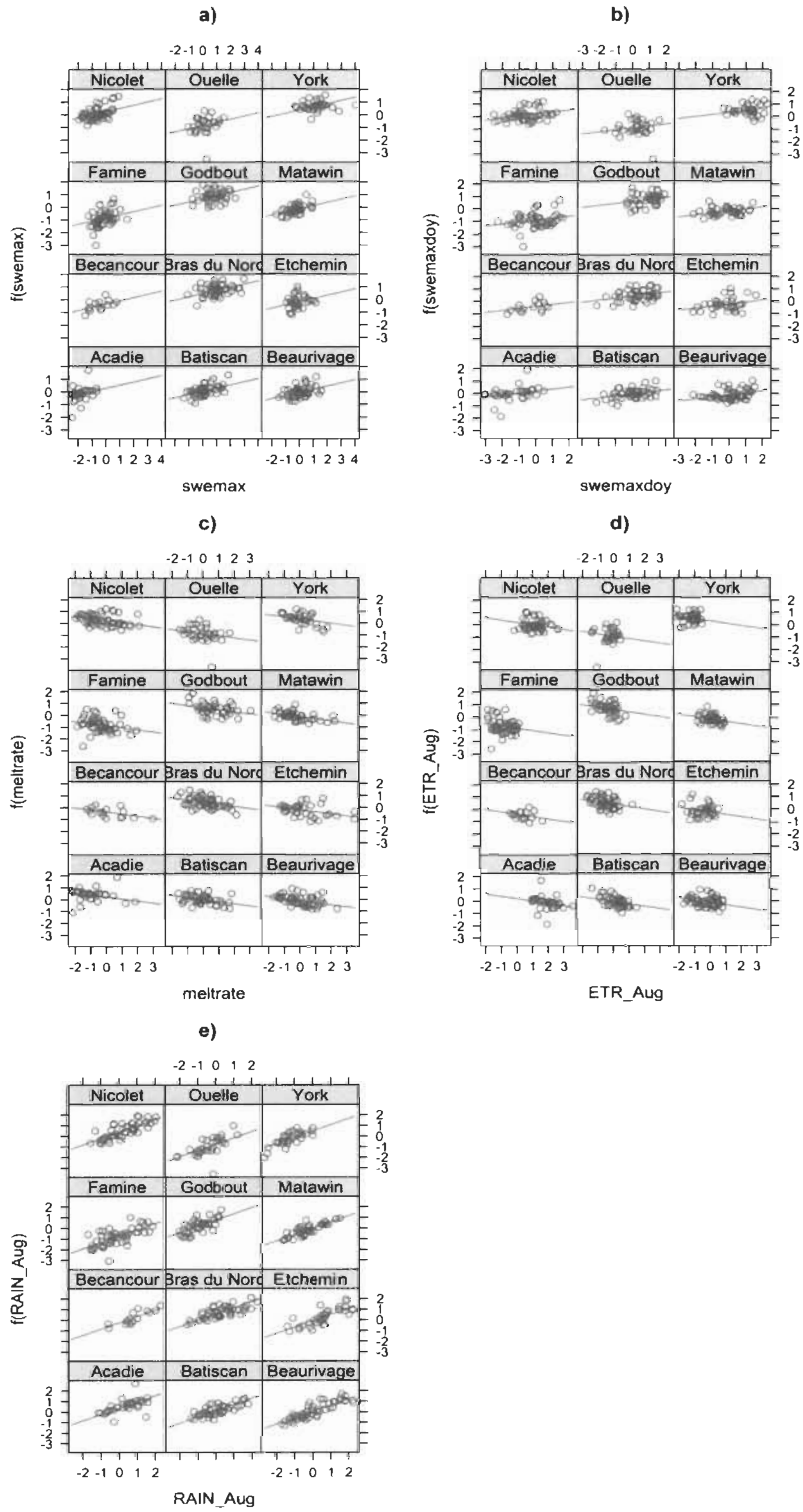

Fig. 5 Partial residual plots of the significant snow-related and summer predictors on summer low flow for the 12 basins separately in August. (a) Maximum SWE; (b) date of occurrence of swemax; (c) melt rate; (d) rainfall; (e) accumulated evapotranspiration.

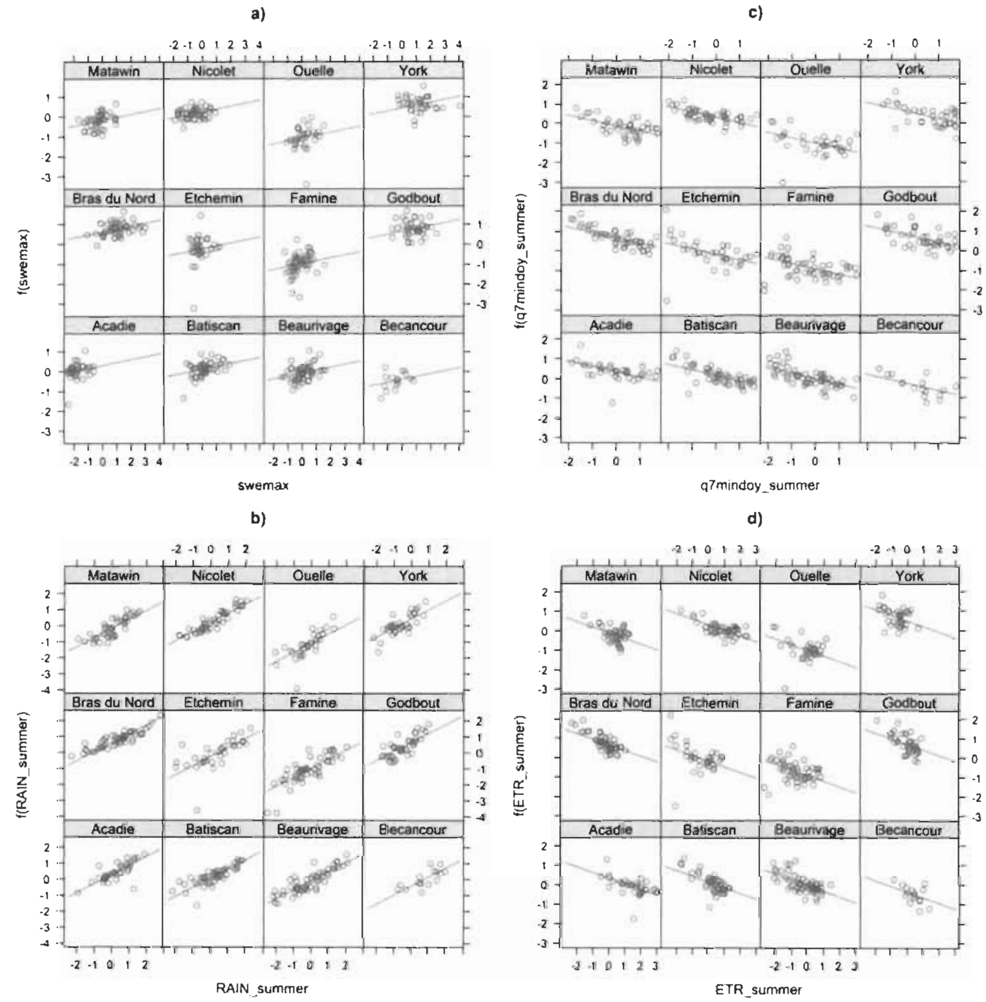

Fig. 6 Partial residual plots of the significant snow-related and summer predictors on summer low flow for the 12 basins separately for the whole summer period. (a) Maximum SWE; (b) low flow occurrence date; (c) rainfall; (d) evapotranspiration.

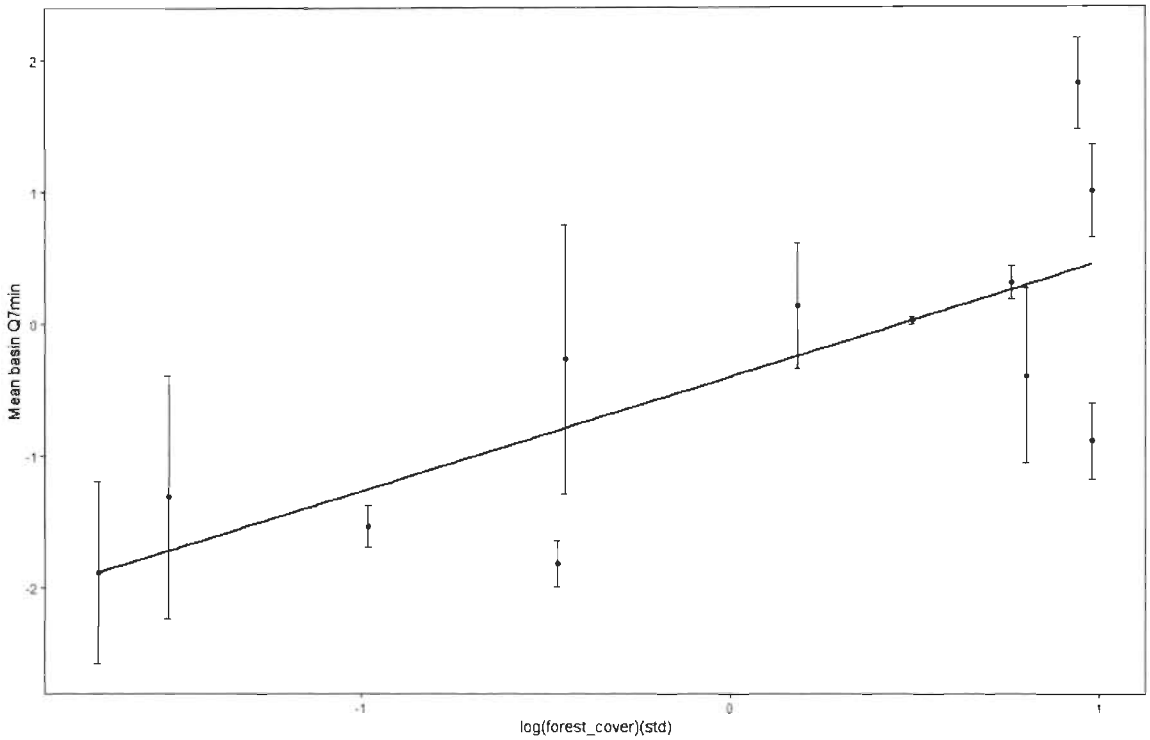

Fig. 7 The sensitivity of minimum discharge to basin forest cover.

Reference

Akaike, H. (1987). Factor analysis and AIC. In Selected papers of hirotugu akaike (pp. 371-386): Springer.

Assani, A. A., Chalifour, A., Légaré, G., Manouane, C.-S., & Leroux, D. (2011). Temporal Regionalization of 7-Day Low Flows in the St. Lawrence Watershed in Quebec (Canada). Water Resources Management, 25(14), 3559-3574. doi:10.1007/s11269-011-9870-6

Assani, A. A., Charron, S., Matteau, M., Mesfioui, M., & Quessy, J.-F. (2010). Temporal variability modes of floods for catchments in the St. Lawrence watershed (Quebec, Canada). Journal of Hydrology, 385(1), 292-299. doi:https://doi.org/10.1016/j.jhydrol.2010.02.031

Aygün, O., Kinnard, C., & Campeau, S. (2020a). Impacts of climate change on the hydrology of northern midlatitude cold regions. Progress in Physical Geography: Earth and Environment, 44(3), 338-375.

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.

Barnett, T. P., Adam, J. C., & Lettenmaier, D. P. (2005). Potential impacts of a warming climate on water availability in snow-dominated regions. Nature, 438(7066), 303-309. doi:10.1038/nature04141

Bavay, M., Lehning, M., Jonas, T., & Löwe, H. (2009). Simulations of future snow cover and discharge in Alpine headwater catchments. Hydrological Processes: An International Journal, 23(1), 95-108.

Beaulieu, M., Schreier, H., & Jost, G. (2012). A shifting hydrological regime: a field investigation of snowmelt runoff processes and their connection to summer base flow, Sunshine Coast, British Columbia. Hydrological Processes, 26(17), 2672-2682.

Bergeron, O. (2015). Grilles climatiques quotidiennes du Programme de Surveillance du Climat du Québec, version 1.2–Guide d'utilisation. Québec, ministère du Développement Durable, de l'Environnement et de la Lutte contre les Changements Climatiques, Direction du suivi de l'état de l'environnement.

Berghuijs, W. R., Woods, R. A., & Hrachowitz, M. (2014). A precipitation shift from snow towards rain leads to a decrease in streamflow. Nature Climate Change, 4(7), 583-586. doi:10.1038/nclimate2246

Blöschl, G., Hall, J., Parajka, J., Perdigão, R. A., Merz, B., Arheimer, B., . . . Borga, M. (2017). Changing climate shifts timing of European floods. Science, 357(6351), 588-590.

Bonsal, B., & Shabbar, A. (2008). Impacts of large-scale circulation variability on low streamflows over Canada: a review. Canadian Water Resources Journal, 33(2), 137-154.

Boumaiza, L., Chesnaux, R., Walter, J., & Stumpp, C. (2020). Assessing groundwater recharge and transpiration in a humid northern region dominated by snowmelt using vadose-zone depth profiles. Hydrogeology Journal, 28(7), 2315-2329.

Boyer, C., Chaumont, D., Chartier, I., & Roy, A. G. (2010). Impact of climate change on the hydrology of St. Lawrence tributaries. Journal of Hydrology, 384(1-2), 65-83.

Burn, D. H., Buttle, J. M., Caissie, D., MacCulloch, G., Spence, C., & Stahl, K. (2008). The processes, patterns and impacts of low flows across Canada. Canadian Water Resources Journal, 33(2), 107-124.

Buttle, J. M., Allen, D. M., Caissie, D., Davison, B., Hayashi, M., Peters, D. L., . . . Whitfield, P. H. (2016). Flood processes in Canada: Regional and special aspects. Canadian Water Resources Journal / Revue canadienne des ressources hydriques, 41(1-2), 7-30. doi:10.1080/07011784.2015.1131629

Comeau, L. E., Pietroniro, A., & Demuth, M. N. (2009). Glacier contribution to the North and South Saskatchewan rivers. Hydrological Processes: An International Journal, 23(18), 2640-2653.

Cooper, M., Schaperow, J., Cooley, S., Alam, S., Smith, L., & Lettenmaier, D. (2018). Climate elasticity of low flows in the maritime western US mountains. Water Resources Research, 54(8), 5602-5619.

Curry, C. L., & Zwiers, F. W. (2018). Examining controls on peak annual streamflow and floods in the Fraser River Basin of British Columbia. Hydrology and Earth System Sciences, 22(4), 2285-2309.

Dierauer, J. R., Whitfield, P. H., & Allen, D. M. (2018). Climate controls on runoff and low flows in mountain catchments of Western North America. Water Resources Research, 54(10), 7495-7510.

Diffenbaugh, N. S., Scherer, M., & Ashfaq, M. (2013). Response of snow-dependent hydrologic extremes to continued global warming. Nature Climate Change, 3(4), 379-384.

Godsey, S. E., Kirchner, J. W., & Tague, C. L. (2014). Effects of changes in winter snowpacks on summer low flows: case studies in the Sierra Nevada, California, USA. Hydrological Processes, 28(19), 5048-5064. doi:10.1002/hyp.9943

Guay, C., Minville, M., & Braun, M. (2015). A global portrait of hydrological changes at the 2050 horizon for the province of Québec. Canadian Water Resources Journal/Revue canadienne des ressources hydriques, 40(3), 285-302.

Hodgkins, G. A., Dudley, R. W., & Huntington, T. G. (2005). Summer low flows in new england during the 20th century 1. JAWRA Journal of the American Water Resources Association, 41(2), 403-411.

Jasechko, S., Birks, S. J., Gleeson, T., Wada, Y., Fawcett, P. J., Sharp, Z. D., . . . Welker, J. M. (2014). The pronounced seasonality of global groundwater recharge. Water Resources Research, 50(11), 8845-8867.

Jenicek, M., Seibert, J., & Staudinger, M. (2018). Modeling of future changes in seasonal snowpack and impacts on summer low flows in alpine catchments. Water Resources Research, 54(1), 538-556.

Jenicek, M., Seibert, J., Zappa, M., Staudinger, M., & Jonas, T. (2016). Importance of maximum snow accumulation for summer low flows in humid catchments. Hydrology and Earth System Sciences, 20(2), 859-874. doi:10.5194/hess-20-859-2016

Laghari, A., Vanham, D., & Rauch, W. (2012). To what extent does climate change result in a shift in Alpine hydrology? A case study in the Austrian Alps. Hydrological Sciences Journal, 57(1), 103-117.

Maurer, E. P., Stewart, I., Bonfils, C., Duffy, P. B., & Cayan, D. (2007). Detection, attribution, and sensitivity of trends toward earlier streamflow in the Sierra Nevada. Journal of Geophysical Research: Atmospheres, 112(D11).

Mazouz, R., Assani, A. A., & Rodríguez, M. A. (2013). Application of redundancy analysis to hydroclimatology: A case study of spring heavy floods in southern Québec (Canada). Journal of Hydrology, 496, 187-194.

Moore, J. N., Harper, J. T., & Greenwood, M. C. (2007). Significance of trends toward earlier snowmelt runoff, Columbia and Missouri Basin headwaters, western United States. Geophysical Research Letters, 34(16).

Mumma, M., Assani, A., Landry, R., Quessy, J., & Mesfioui, M. (2001). Effect of the change from forest to agriculture land use on the spatial variability of summer extreme daily flow characteristics in Southern Quebec (Canada). Journal of Hydrology, 407(1-4), 153-163.

Musselman, K. N., Clark, M. P., Liu, C., Ikeda, K., & Rasmussen, R. (2017). Slower snowmelt in a warmer world. Nature Climate Change, 7(3), 214-219.

Neary, D. G., Ice, G. G., & Jackson, C. R. (2009). Linkages between forest soils and water quality and quantity. Forest ecology and management, 258(10), 2269-2281.

Nemri, S., & Kinnard, C. (2020). Comparing calibration strategies of a conceptual snow hydrology model and their impact on model performance and parameter identifiability. Journal of Hydrology, 582, 124474.

O'brien, R. M. (2007). A caution regarding rules of thumb for variance inflation factors. Quality & quantity, 41(5), 673-690.

Perrin, C., Michel, C., & Andréassian, V. (2003). Improvement of a parsimonious model for streamflow simulation. Journal of Hydrology, 279(1-4), 275-289.

Pike, R., & Scherer, R. (2003). Overview of the potential effects of forest management on low flows in snowmelt-dominated hydrologic regimes. Journal of Ecosystems and Management, 3(1).

Pomeroy, J. W., Fang, X., & Marks, D. G. (2016). The cold rain-on-snow event of June 2013 in the Canadian Rockies—Characteristics and diagnosis. Hydrological Processes, 30(17), 2899-2914.

Quilbé, R., Rousseau, A., Moquet, J.-S., Savary, S., Ricard, S., & Garbouj, M. (2008). Hydrological responses of a watershed to historical land use evolution and future land use scenarios under climate change conditions. Hydrology and Earth System Sciences, 12, 101-110. doi:10.5194/hessd-4-1337-2007

Riggs, H. C. (1978). Streamflow characteristics from channel size. Journal of the Hydraulics Division, 104 (ASCE 13501 Proceeding).

Sakia, R. M. (1992). The Box-Cox transformation technique: a review. Journal of the Royal Statistical Society: Series D (The Statistician), 41(2), 169-178.

Schaefli, B., Hingray, B., Niggli, M., & Musy, A. (2005). A conceptual glacio-hydrological model for high mountainous catchments. Hydrology and Earth System Sciences, 9. doi:10.5194/hessd-2-73-2005

Smakhtin, V. U. (2001). Low flow hydrology: a review. Journal of Hydrology, 240(3), 147-186. doi:https://doi.org/10.1016/S0022-1694(00)00340-1

Stewart, I. T., Cayan, D. R., & Dettinger, M. D. (2004). Changes in snowmelt runoff timing in western North America under a business as usual climate change scenario. Climatic Change, 62(1-3), 217-232.

Sturm, M., Goldstein, M. A., & Parr, C. (2017). Water and life from snow: A trillion dollar science question. Water Resources Research, 53(5), 3534-3544.

Tian, P., Zhao, G.-j., Li, J., & Tian, K. (2011). Extreme value analysis of streamflow time series in Poyang Lake Basin, China. Water Science and Engineering, 4(2), 121-132.

Valéry, A., Andréassian, V., & Perrin, C. (2014). 'As simple as possible but not simpler': What is useful in a temperature-based snow-accounting routine? Part 2–Sensitivity analysis of the Cemaneige snow accounting routine on 380 catchments. Journal of Hydrology, 517, 1176-1187.

Veijalainen, N., Lotsari, E., Alho, P., Vehviläinen, B., & Käyhkö, J. (2010). National scale assessment of climate change impacts on flooding in Finland. Journal of Hydrology, 391(3-4), 333-350.

Yang, D., Kane, D. L., Hinzman, L. D., Zhang, X., Zhang, T., & Ye, H. (2002). Siberian Lena River hydrologic regime and recent change. Journal of Geophysical Research: Atmospheres, 107(D23), ACL 14-11-ACL 14-10.

Zhang, F.-Y., Li, L.-H., Ahmad, S., & Li, X.-M. (2014). Using path analysis to identify the influence of climatic factors on spring peak flow dominated by snowmelt in an alpine watershed. Journal of Mountain Science, 11(4), 990-1000. doi:10.1007/s11629-013-2789-z

CHAPITRE III

IMPACT OF SNOW COVER ON LOW FLOW IN NORTH AMERICA

Article en attente de soumission au journal scientifique *Environnemental Research Letters*

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Abstract

Limited information exists on the linkages between snowmelt and summer low flow generation in cold regions. The objective of this research was to demonstrate the importance of snow cover on low flow during the warm season for a sample of 260 snow-affected pristine river basins in North America. Correlations between summer low flow magnitude and antecedent winter and summer climate variables showed that the maximum winter snow water equivalent (SWE) influenced summer low flow, with a decreasing influence of SWE throughout summer and significant inter-basin heterogeneity. The sensitivity of low flow to both SWE and summer rainfall was assessed using multivariate hierarchical models which include the effects of basin characteristics. The sensitivity of low flows to SWE was more important in high-elevation basins and decrease for low-elevation ones specially in July and August and less important in forested basins overall in summer. On the other hand, the results show that the effect of maximum SWE on low flow is decreasing towards September. The snowmelt contributes more the minimum discharge in July (median rho in July = 0.21). However, snow accumulation could only partly explain to the observed low flow inter-annual variations. One other important factor was the precipitation between maximum SWE and summer low flows. As expected in most of the selected basins a humid summer can lead to a higher low flow magnitude. We assessed the sensitivity of individual catchments to the change of maximum SWE indicating a decreasing of sensitivity of summer low flow to snow accumulation at the end of summer compared to July. The results of hierarchical model show and important role of basins descriptors on low flows, that the magnitude of low flows are higher in forest basin (α= 0.31-33) than in agricultural one in July and September however the low flows volume is more important in lower altitude basins and steeper slope in August.

Keywords: Climate change; low flows; interannual variability, occurrence date, magnitude, snow water equivalent, hydrology, catchment descriptors.

Introduction

Low flows are a seasonal phenomenon defined as the "lowest annual level reached by a river at a given point" (Roche, 1986), and is generally characterized by the minimum annual flow or the minimum of average flow over 7 or 30 days. It is a natural event resulting from deficits in precipitation during an extended period. Low flow in summer is important maintaining riparian ecosystems as it determines the amount of habitat available for aquatic species (Smakhtin, 2001). Low flows influence the concentration of pollutants in river systems, as prolonged hydrological droughts lead to an artificial increase in dissolved chemical concentrations which can kill certain aquatic species or populations (Anctil et al., 2000; Burn et al., 2008). On the other hand, minimum discharge in summer is also important for river infrastructure, reservoir storage design, river transport regulation, supplying potable water, navigation and for maintaining the quality and quantity of water for irrigation (Assani et al., 2005; Godsey et al., 2014).

Hydrological drought severity is highly dependent on terrestrial hydrological processes, which are governed by a combination of climate and catchment controls (Burn et al., 2008; Van Loon et al., 2015). Several processes can drive interannual variability of low flows. Summer low flows can be conditioned by antecedent summer meteorological conditions, particularly the interplay between rainfall and evapotranspiration, but also by the slow release of water stored in aquifers, lakes, and glaciers (Freeze, 1974). The gradual depletion of discharge during periods with little or no precipitation thus depends on the storage capacities and rates of water transmission (Tallaksen, 1995), which are themselves influenced by basin characteristics such as soil texture, depth of surficial deposits and the presence of fractured bedrocks, but also by seasonal storage of water in snowpacks (Van Loon et al., 2015). Land cover also modulates low flows hydrology. Several studies, most carried out in basins not affected by snow, have demonstrated that an increase in basin forest cover is related to lower low flows volumes attributed to higher evapotranspiration rates in forests, while other studies have reported increasing low flows volume with higher basin forest cover, attributed to higher infiltration and recharge of subsurface storage (Price, 2011). Understanding the relationship between soil moisture infiltration capacities and precipitation is thus necessary to understand low flows generation processes (Jenicek et al., 2016).

Cold region hydrology is shaped by the accumulation of snow in winter and its melt during spring, which impacts seasonal water resource availability (Barnett et al., 2005). Melting of the winter snowpack releases large volumes of water which causes springtime flooding in snow-dominated catchments (Pomeroy et al., 2013), but also replenish aquifers (Stewart et al., 2004). Snowmelt is known to be a more efficient contributor to recharge than rainfall because snowmelt occurs over a prolonged period with minimal evapotranspiration losses and at a lower intensity compared to rainfall events, which occur during the vegetation growing season and are often of high intensity and short duration, promoting infiltration excess runoff (Earman et al., 2006). Hence annual groundwater recharge can be more sensitive to winter recharge processes than to summer recharge in arid and temperate climates (Jasechko et al., 2016). This could explain why basins with an important fraction of snowfall in winter are associated with higher long-term mean streamflow (Berghuijs et al., 2014; Jenicek et al., 2020). However, the snowmelt contribution to ground water recharge is affected by many factors, such as radiation, topography and vegetation which affect melting rates and thus infiltration (Pomeroy et al., 2012). Thus, knowledge of snow conditions in winter and spring can help to predict in advance water availability in the following summer.

Most previous empirical studies investigating hydroclimate controls on low flows have linked interannual variability of minimum summer discharge to cumulative precipitation or temperature in summer (Godsey et al., 2014; Hodgkins et al., 2005; Jenicek et al., 2016; Dierauer et al., 2018; Cooper et al., 2018). Comparatively fewer studies have investigated the linkages between summer low flows regimes and antecedent snow storage. Some studies have looked at the modes of interannual variability of low flows in northern regions and their link to snowpack at the annual scale (Beaulieu et al., 2012; Van Loon et al., 2015). Regional studies in humid mountain catchments such as the Swiss and Austrian Alps (Jenicek et al., 2016; Jenicek et al., 2020; Laghari et al., 2012) and in semiarid western U.S. mountains (Cooper et al 2018; Dierauer et al., 2018; Godsey et al. 2014), have shown that snow conditions in winter can affect summer low flows, particularly in areas where winter and summer precipitation are highly variable, since the total amount of precipitation accumulated in winter and melted in spring affects groundwater recharge and thus river flow during dry summer periods. Among these studies, Godsey et al. (2014) found from historical records in eight watersheds in the Sierra Nevada, western USA, that a decrease in maximum snow water equivalent (SWE) of 10% resulted in a decrease in minimum summer discharge of 9-22% and an earlier occurrence of low flows by 3-7 days. Similar results were reported in alpine and subalpine watersheds in Switzerland, where maximum SWE showed the best predictive ability for 7-day summer low-flow among all winter-related predictors (Jenicek et al., 2016). Jenicek et al. (2016) also showed that summer low flows were most sensitive to SWE in high-elevation basins, and that the date of occurrence of the summer minimum discharge was positively correlated with maximum SWE in medium to high elevation (> 1500 m) catchments. This may be related to the large amount of available water released by snow in the high-elevation basins, despite the steeper slopes and shallow soils which promote rapid water transfers downstream (Staudinger et al., 2017). In maritime mountains catchments in western U.S, Cooper et al.(2018) showed that the sensitivity of low flows to SWE and evapotranspiration depends on the climate characteristics in different sites and drainage density. These studies suggest that, the response of low flows to SWE and evapotranspiration demand variability is larger in semiarid climate than in the humid Pacific Northwest in California, and the sensitivity of low flows to climate change is less important in slow-draining basins.

Overall, these previous studies showed the importance of snowfall and maximum SWE on summer low flows in arid (Sierra Nevada, western US) and temperate (Swiss Alps) mountain basins. However, these regional studies mostly focused on mountainous catchment, and the small number of catchment studied impedes the generalisation of their findings to other types of basins and climates. In this study, the relative impact of snow storage and summer climate on summer low flows were assessed statistically for a large sample of basins in North America, representing various climatic, topographic and land use conditions. We seek to answer a simple question: what are the respective contributions of snowmelt and summer precipitation to the magnitude of low flows events in summer? Do basin biophysical conditions have a damping or an amplifying impact on the sensitivity of low flows to snow cover?

Methodology

Description of study sites

The ratio between snowfall and surface runoff (equation 1, Barnett et al., 2005) was used to define regions of the NH where snowmelt plays a dominant role in seasonal runoff.

$$R = \frac{Snowfall}{Surface\ runoff} \tag{1}$$

The ratio was calculated using average surface runoff and snowfall over the 1980-2018 period from MERRA-2 reanalysis product (Gelaro et al., 2017). Grid cells with $R > 0.5$ where flagged as being significantly influenced by snowmelt, following Barnett et al. (2005). Daily historical streamflow observations from gauging stations within this region were then obtained from the Climate Sensitive Stations Dataset (Pristine River Basins) of the Global Runoff Data Center (GRDC) (Grabs et al., 1996), which is a global archive of river discharge data with natural or near-natural hydrological regimes. Watershed limits for the selected stations based on the HydroSHEDS product (Lehner et al., 2008) were obtained from the GRDC. Streamflow records were screened according to the following criteria: the selected stations had to have near-continuous daily records during the entire study period from 1958 to 2018, i.e. stations with more than 10% of missing data in any single year, and with more than 20% missing year data over the whole period, were excluded. As such, twenty-three stations were removed due to large gaps in their streamflow records. As the study focuses on the impact of seasonal snow cover on low flows conditions, basins containing glaciers were eliminated from the analysis, based on the global Randolph glacier Inventory (RGI) version 60 (Pfeffer et al., 2014). As only thirteen basins were found in Europe, they were excluded to avoid a continental bias in the analysis, and the study thus focus on North America. A few very large basins (area > 50,000 km^2) were also removed to avoid a scale bias in this study. The final dataset includes 260 pristine river basins in North America, i.e. whose flow is largely unaffected by anthropogenic activities (Fig. 1).

Climate and biophysical datasets

To analyse statistical relationships between snow accumulation, summer climate conditions (precipitation, evapotranspiration) and observed streamflow, long records with high spatial and temporal resolutions are desirable. Monthly basin averages of SWE, precipitation and evapotranspiration were extracted from TerraClimate, a high-resolution (1/24, ~4-km) global dataset of monthly climate and climatic water balance from 1958-2015 (Abatzoglou et al., 2018). TerraClimate uses climatically aided interpolation, combining high-spatial resolution climatological normals from the WorldClim dataset, with coarser spatial resolution, but time-varying data from CRU Ts4.0 and the Japanese 55-year Reanalysis (JRA55). The use of a global gridded hydroclimate datasets circumvents using observational records which suffer from frequent gaps and need to be interpolated to the whole basin. TerraClimate was preferred over global reanalysis products such as MERRA2 (Gelaro et al., 2017) and ERA5 (Hersbach et al., 2020), which while having sub daily temporal resolution, had shorter historical periods (since 1979 and 1950, respectively) and coarser spatial resolutions (~50 km and ~31 km, respectively) that were larger than the area of roughly half of the studied basins (Fig. 1). TerraClimate was thus the best compromise between record length and spatial and temporal resolution, produces monthly data using surface water balance model that combines evapotranspiration, precipitation, temperature, and plant-extractable water capacity of soil. Its availability on the Google Earth Engine platform (Gorelik et al., 2017) further facilitated the data extraction. Several basin biophysical descriptors were also considered. Mean basin slope (°) and elevation (m) were calculated from the Global Multi-resolution Terrain Elevation Data 2010 (GMTED2010, 7.5 arc-seconds, ~230 m) provided by the U.S. Geological Survey (USGS), which has been widely used for a variety of hydrological, climatological and geomorphological applications (Danielson et al., 2011). Forest cover (%) in each basin was extracted from the Moderate Resolution

Imaging Spectroradiometer (MODIS) land cover type data dataset MCD12Q1 version 6, which is derived from supervised classifications of MODIS Terra and Aqua reflectance data (Sulla-Menashe et al., 2018). This product is widely used globally and regionally and provides 500 m spatial resolution global land cover types at annual intervals (2001-2016). The dominant class of the Annual International Geosphere-Biosphere Programme (IGBP) classification over the 2001-2016 period was used at each grid point. All forest types (evergreen, deciduous, mixed), as well as closed shrublands dominated by woody perennials were considered to calculate the forest cover. The statistical analysis is based on a common 61 years period, from 1958-2018.

Antecedent hydrological conditions and statistical analysis of summer low flows

To analyze climate controls on the low flows regimes, the magnitude of low flows (Q7min) during each summer month (July, August and September) was used as response variable and compared against seven antecedent climate predictor variables derived from TerraClimate (Table 1). Q7min was calculated as the minimum of 7-day moving average of flow during in each summer month. Predictors related to snow cover include the maximum SWE (swemax) between January and May and its date of occurrence (swemaxdoy). The snow disappearance date (meltoutdoy) was defined as the first month with minimum SWE following the maximum SWE date. Summer predictors include the accumulated actual evapotranspiration (ETR) and liquid precipitation (RAIN) between (and including) June and August. Since only total precipitation was available in TerraClimate, this period was used to reduce the possibility of including snowfall within summer rainfall estimates. Rainfall and ETR were accumulated from June up to the month of interest and used as summer climate predictors in each model.

A bivariate correlation analyse was first applied between Q7min and individual predictors to assess the strength and directions of their relationships. The analysis was carried out globally with all basins pooled together and then separately for each basin, to estimate the changing importance of snow contribution to low flows. Then, multivariate hierarchical, or 'multilevel' regression models were used to quantify the respective impact of winter and summer climate variables on the interannual variability of summer low flows, while also considering the effect of basins properties. Hierarchical regression is a widely used in hydrology, statistical model used when observations are organized (grouped) at more than one level and that allows to jointly model the within- and between-group variability (Gelman et al., 2006). The model relates the response variable (Q7min) to hydroclimatic predictors (first level fixed effects) with basins as grouping variable (random effect) and basin descriptors as group-level variables (second level fixed effects) (equation 1):

$$y_t = \alpha_{j[t]} + \beta_{j[t]} x_t + \varepsilon_t, \text{ for } t = 1,\ldots, n \tag{1a}$$

$$\alpha_j = \gamma_0^{\alpha} + \gamma_1^{\alpha} u_j + \eta_j^{\alpha}, \text{ for } j = 1,\ldots, J \tag{1b}$$

$$\beta_j = \gamma_0^{\beta} + \gamma_1^{\beta} u_j + \eta_j^{\beta}, \text{ for } j = 1,\ldots, J \tag{1c}$$

where x_t represent annual-level hydroclimate predictors (n = 61 years) and u_j biophysical descriptors at the basin level (J = 260 basins). $\alpha_{j[t]}$ and $\beta_{j[t]}$ are respectively the intercept and regression slopes of hydroclimatic predictors (first-level fixed effects), which vary between basins as a function of basins descriptors (equation 1b and c). η_j^{α} and η_j^{β} are respectively the between-basin random errors on the intercept and slopes and ε_t the within-basin residual error (Gelman et al., 2006). A top-down strategy was used for model structure selection (Zuur et al., 2009). Starting from the most complex (i.e. the 'beyond optimal', Zuur et al, 2009) model the Akaike Information Criteria (AIC) was first used to choose the optimal structure of the random components, and then fixed components (predictors) were successively removed based on t-tests ($p < 0.05$). Variables deviating from a normal distribution (Q7min, Q7mindoy, Area, medianslope, medianelevation, forestfraction, latitude) were transformed with a BoxCox transformation (Sakia, 1992). A global mean-centering and standardization (Z score) was applied on all variables to use a common scale in the analysis, while preserving inter-basin differences. As such, the standardized partial regression slopes represent the sensitivity of Q7min to each predictor while holding other predictors constant.

Results and discussion

Inter-basin variability in snow disappearance and low discharge

There is considerable diversity of hydroclimatic conditions across the 260 basins (Fig. 1). The snow disappears earlier at lower elevations and low latitudes, generally in February and March and persists longer at higher elevations and latitudes, sometimes up to August in alpine basins (Supplementary Fig. 2). The occurrence date of low flows also varies among basins but is more homogeneous (Fig. 1), occurring on average between July and September with the earliest values in low altitude alpine basin (> 1000 m, Supplementary Fig. 2).

Correlation of low flows with hydroclimate predictors

The correlation between summer low flows (Q7min) and the antecedent hydroclimate variables show a clear seasonal pattern (Fig. 3). Generally, Q7min is positively correlated with swemax and swemaxdoy ($p < 0.05$), with maximum positive mean correlation occurring in July and the median correlation decreasing over time, from 0.21 in July to 0.19 in September. This suggests that larger and later snowpacks tend to favor higher low flows in summer, but with a decreasing influence over time. As expected, stronger relationships are seen between summer predictors and Q7min. The accumulated rainfall (RAIN) shows the strongest correlations, peaking in July (median rho = 0.61) and decreasing over time (median rho in September = 0.56). A similar pattern is seen for accumulated evapotranspiration (ETR), albeit with reduced correlations compared to RAIN. These results suggest that wetter summers lead to high low flows, as expected, but that higher evapotranspiration losses are also associated with higher low flows, which is counterintuitive. This is because ETR is strongly correlated (rho = 0.87) with rainfall (Fig. 3). This occurs because most of the studied basins have a water-limited regime (Fig. 1e), so that interannual variations in ETP are mostly driven by water availability. No clear correlation patterns emerged between Q7min and its timing (Q7mindoy), so that globally, low flows of lesser magnitude do not necessarily occur later from one year to the other. The boxplots in Fig. 2 show large inter-basins variability in the correlation.

We hypothesize that this heterogeneity reflects a legacy effect of the basin characteristics on the relationships between hydroclimatic variables and Q7min, which is examined further.

The spatial heterogeneity in the correlations of Q7min with maximum SWE and summer rainfall is best seen in Fig. 3. The correlation between swemax and Q7min is strongest in high elevation basins of western North America and decreases in the prairies and towards the eastern regions. This may be connected to thicker snowpacks in high elevation basin and the slower melting in spring which promote meltwater infiltration (Jenicek et al., 2016). The correlation between swemax and Q7min reaches low positive or negatives values in the southern fringe of Eastern North America but also further North in August and September. While there is an overall small decrease in correlation over time (Fig. 2), the correlation patterns remain essentially the same between July and September (Fig. 3a-c).

The stronger correlation between summer low flows and summer rainfall shows a spatial pattern somewhat inverse to that for SWE: correlations are greater in basins with a humid continental climate, i.e. in central-eastern North America and weaker in the alpine and semiarid western region (Fig. 3d-f). Hence a broad pattern emerge which reflects the interplay between elevation and climate. Western mountain basins with thicker snowpacks and a semiarid climate with limited summer precipitation favor a larger contribution of snowmelt to summer low flows. Conversely, low flows variability is more related to rainfall variability in lower elevation basins, where snow storage is reduced but summer rainfall more abundant. However, there is significant scatter in this broad pattern which suggests that other catchment properties and climate conditions may play a role in this heterogeneity (Burn et al., 2008). To disentangle the respective effects of snow storage relative to that of rainfall, a multivariate approach is needed and explored in the next section.

Multivariate analysis

A strong collinearity between ETR and RAIN was identified (Supplementary Fig. 2), which arises due to the aforementioned water-limited evapotranspiration regime. A strong correlation between swemax and swemean (Supplementary Fig. 2) and between forestfraction and medianslope (rho = 0.65) was also found. So, ETR, swemean and medianslope were thus excluded from the multivariate hierarchical models.

Results from the monthly hierarchical models are presented in Table 2. Fixed effects, i.e. first-level hydroclimate variables and second-level basin descriptors, explain 35-48% of the interannual variability in Q7min (marginal R^2 = 0.35-0.48, Nakagawa et al., 2013). The difference between marginal and conditional R^2 (conditional R^2 = 0.80-0.83) represent significant inter basin variability (34-44%) in Q7min not explained by fixed factors. Among the hydroclimate (1st level) variables, the maximum SWE (swemax) shows a positive effect on Q7min, decreasing over the course of summer, in agreement with the bivariate correlation result (Fig. 2 and 3). The occurrence date of maximum SWE and the snow disappearance dates have only a marginal effect on Q7min. Summer rainfall (RAIN) has the strongest effect on Q7min, also in agreement with the correlation results. However, and different to the simple bivariate relationships, the relative importance of SWE, as estimated by its partial slope, is higher when other predictors are considered. As such the sensitivity of Q7min to swemax relative to RAIN is 55%, 50% and 46% in July, August and September, respectively. The low flows timing (Q7mindoy) is found to have a small negative influence on low flows magnitude, so that earlier low flows tend to have a larger magnitude, consistent with that fact that a longer period of precipitation deficit leads to a progressive drawdown of aquifers and decreasing river baseflow. This is especially pronounced in September (Table 2), when autumn rainfall typically interrupts the summer drought conditions during which groundwater supports baseflow (Tague et al., 2009).

Among the 2nd level variables (basin descriptors), forest cover is seen to play an important role in increasing the magnitude of Q7min in July, August and September. We interpret this to reflect the higher infiltration capacity of forest soils due to porous organic layer and root networks, compared to agricultural fields with frequently compacted soils and often clay-rich soils, which restrict infiltration (Neary et al., 2009). These results are in agreement with studies conducted in North America that showed that a higher minimum extreme flow was found in the forested watersheds comparing to agricultural watersheds (Caissie et al., 2002; Pike et al., 2003; Mumma et al., 2001; Quilbé et al., 2008; Sylvain et al., 2015). Latitude has a consistent positive effect on Q7min throughout summer, so that northern basins tend to have higher low flows. We interpret this to mainly result from the reduced evapotranspiration losses towards higher latitudes (Yuan et al., 2010).

As shown in Fig. 4, there is some variability, but no clear spatial pattern of Q7min sensitivities to swemax and RAIN. The patterns in Fig. 4a-c reflect the decreasing sensitivity of Q7min to swemax from July to September, while the pattern for rainfall is more constant over time. Factors explaining these patterns can be sought in the cross-level interactions, i.e. how basin descriptors modulate the low flow response to SWE and rainfall (Table 2). While basin elevation hasn't any effect on low flows, a significant interaction was found between the swemax sensitivity and elevation in July-August, with Q7 tending to be more sensitive to SWE in higher basins than in lowlands, which probably reflects the larger snow storage at higher elevations which lead to more efficient recharge and subsurface water transfers (Birsan et al., 2005; Barnhart et al., 2016; Earman et al., 2006; Staudinger et al., 2017; Jasecho et al., 2016). A similar result was found in the Alpine and pre-Alpine catchments in Switzerland, like (Jenicek et al., 2016) a positive significant effect of basin elevation on the sensitivity of Q7min to swemax was found. A small positive effect of latitude was found on the sensitivity of Q7min to swemax in July-August, perhaps reflecting the reduced evapotranspiration losses of snow melt water in spring at higher latitudes. A negative interaction is found between maximum SWE and forest cover, suggesting that Q7min is less (more) sensitive to swemax in forested (open) terrains. This could be explained by the snow interception effect, whereas a fraction of snowfall is stored on the canopy and lost by sublimation (Sun et al., 2018). Since this process is not explicitly represented in TerraClimate, overestimated swemax in forests could lead to a decreased sensitivity of Q7min to swemax. A small and negative effect of elevation on the sensitivity of Q7min to rainfall was also found in July and September, suggesting a reduced influence of rainfall on Q7min variability in higher elevation basins, i.e. the inverse relationship than for swemax.

There are still some uncertainties that affect our results, mainly from the global data source. The limitations come from the uncertainties in data quality as well as the analysis method. Despite TerraClimate allows for a high-resolution as well as time period covered dataset some uncertainties exist in the estimation of meteorological datasets (Morice et al., 2012). For example, measurement and sampling uncertainties in TerraClimate are a combination of uncorrelated measurement (New et al., 2000) and climate inhomogeneities that contribute to likely uncertain data temperature, precipitation, and vapor pressure fields, as well as estimated ET0. Moreover, TerraClimate use a simple water balance model based only on a combination of the essential climate variables and does not account for heterogeneity in vegetation types and the impact of changing environmental conditions on their physiology which lower the predictive ability of the hierarchical model and make the predicted low flows potentially uncertain. Overall TerraClimate estimated microclimate features represents likely errors in complex terrain or heterogeneous land-cover. Therefore, using empirically based models for daily precipitation, minimum and maximum temperature in conjunction with regional reanalysis data, modelled daily soil moisture, along with topographic factors and canopy cover might be a good way to improve our understanding of the low flows mechanisms in the northern countries.

Conclusions

In cold regions, low flows are of paramount important for ecological integrity and human socio-economic welfare. In this study we examined and quantified the influence of winter snow conditions on summer low flow for a large sample of North American basins for the period 1958 to 2018. Overall, winter snow storage as estimated with the maximum SWE significantly affected the interannual variability of low flows in summer from July to September. Thus, an important snow accumulation in winter is generally related to a higher low flow magnitude later in the year, but the results showed heterogeneity between basins. Stronger bivariate correlations between swemax and Q7min were found in semi-arid mountains basins of western North America, while Q7min was better correlated with rainfall in the central-eastern continental-humid areas. However, when considering the joint influence of snow conditions and summer rainfall, the sensitivity of Q7min to swemax was almost always positive and amounted to ~50% of the sensitivity to rainfall. The spatial pattern of swemax sensitivities is more complex but was found to be partly related to latitude and altitude, both of which increase the sensitivity of Q7min to SWE, and to forest cover which decreased it. The results of this study provide another perspective of the impacts of climate change on snowmelt. The reduced snow storage anticipated in the mid latitudes (Aygun et al., 2020a) and mountainous regions of North America (Barnett et al., 2005; Mankin et al., 2015) could thus exacerbate hydrological droughts in summer and lead to ecosystem degradation and water quality and scarcity issues.

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Table 1. Hydroclimate variables used in this study.

Variables	Abbreviation (Units)
Predictor variables	
Maximum SWE between January and May	swemax (mm)
Occurrence date of maximum of SWE	Swemaxdoy (day of year)
Snow disappearance date	Meltoutdoy (month)
Sum of summer precipitation from June to August	RAIN (mm)
occurrence date of 7-day minimum discharge	Q7mindoy (day of year)
Sum of actual evapotranspiration from June to August	ETR (mm)
Response variable	
Minimum of 7-day moving average of discharge	Q7min (mm)

Table 2. Results of hierarchical models of low flow magnitude against the six antecedent hydroclimate predictors and basins descriptors for the 260 basins. '-' represent non-significant coefficients ($p < 0.05$).

	Q7min July	**Q7min August**	**Q7min Sept.**
Predictors	Estimates	Estimates	Estimates
(Intercept)	-	-	-
1st level (hydroclimate) predictors			
swemax	0.22	0.19	0.14
swemaxdoy	-	-0.02	-0.02
meltoutdoy	-0.01	-0.01	-
RAIN	0.33	0.31	0.32
Q7mindoy	-0.01	-0.01	-0.05
2nd level (basin descriptors)			
Area	-	-	-
Latitude	0.22	0.20	0.21
Elevation	-	-	-
Forestfraction	0.31	0.32	0.33
Cross-level interactions			
RAIN * Area	-	-	-
RAIN * Latitude	0.03	-	0.03
RAIN* forestfraction	-0.05	-0.3	-
RAIN * medianelevation	-0.05	-	-0.06
swemax *Area	-	-	-
swemax* Latitude	0.08	0.06	-
swemax * forestfraction	-0.12	-0.10	-0.08
swemax *medianelevation	0.10	0.08	-
Random Effects			
τ_{00}	0.33	0.29	0.38
Marginal R^2 / Conditional R^2	0.431 / 0.830	0.480 / 0.822	0.354 / 0.798

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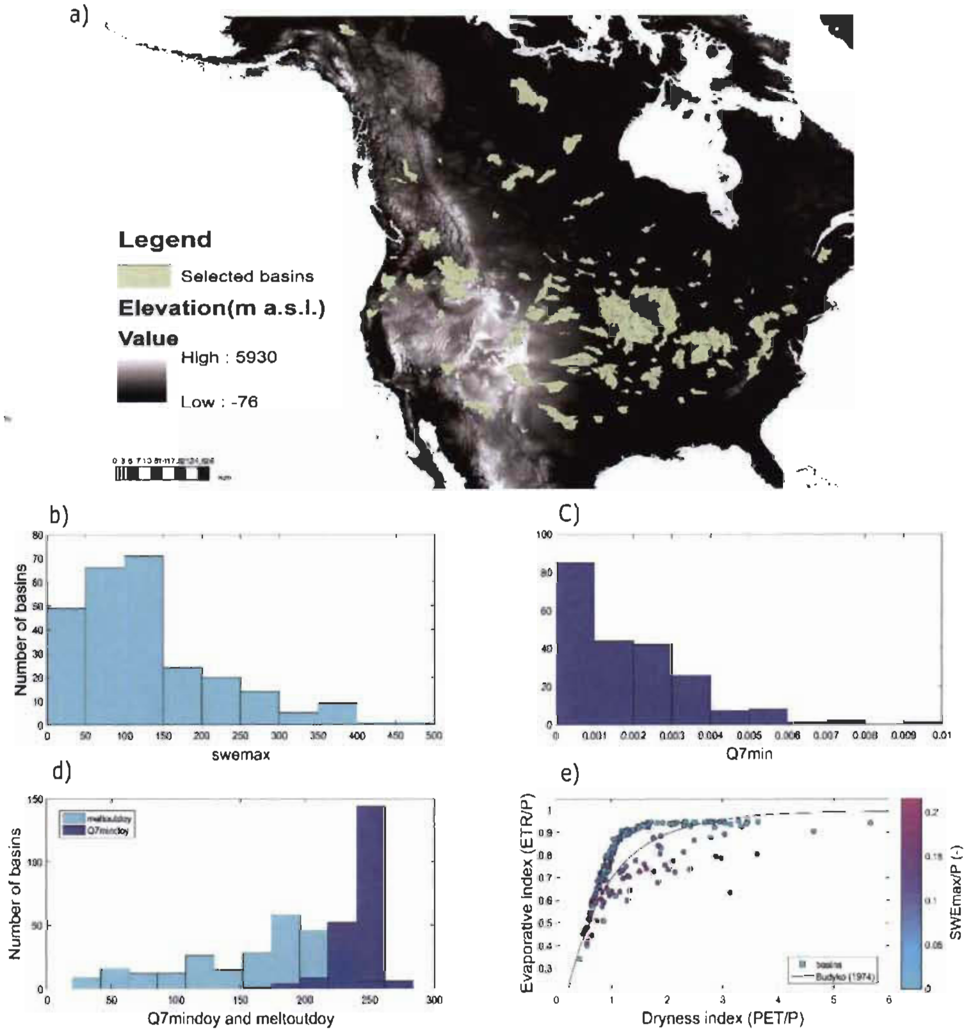


Fig. 1 Location map. (a) selected basins (green) in North America; (b) Mean maximum SWE per basin; (c) mean timing of minimum summer low flow (Q7mindoy) and snow disappearance date (swemaxdoy); (d) mean basin minimum summer low flow (Q7min); (e) Budyko plot of the selected basins, with colour scale showing the mean net annual precipitation, and the red and blue lines the energy and water limit, respectively. Dark blue colours represent low flow related variables and light blue snow related variables. In panels b-d.

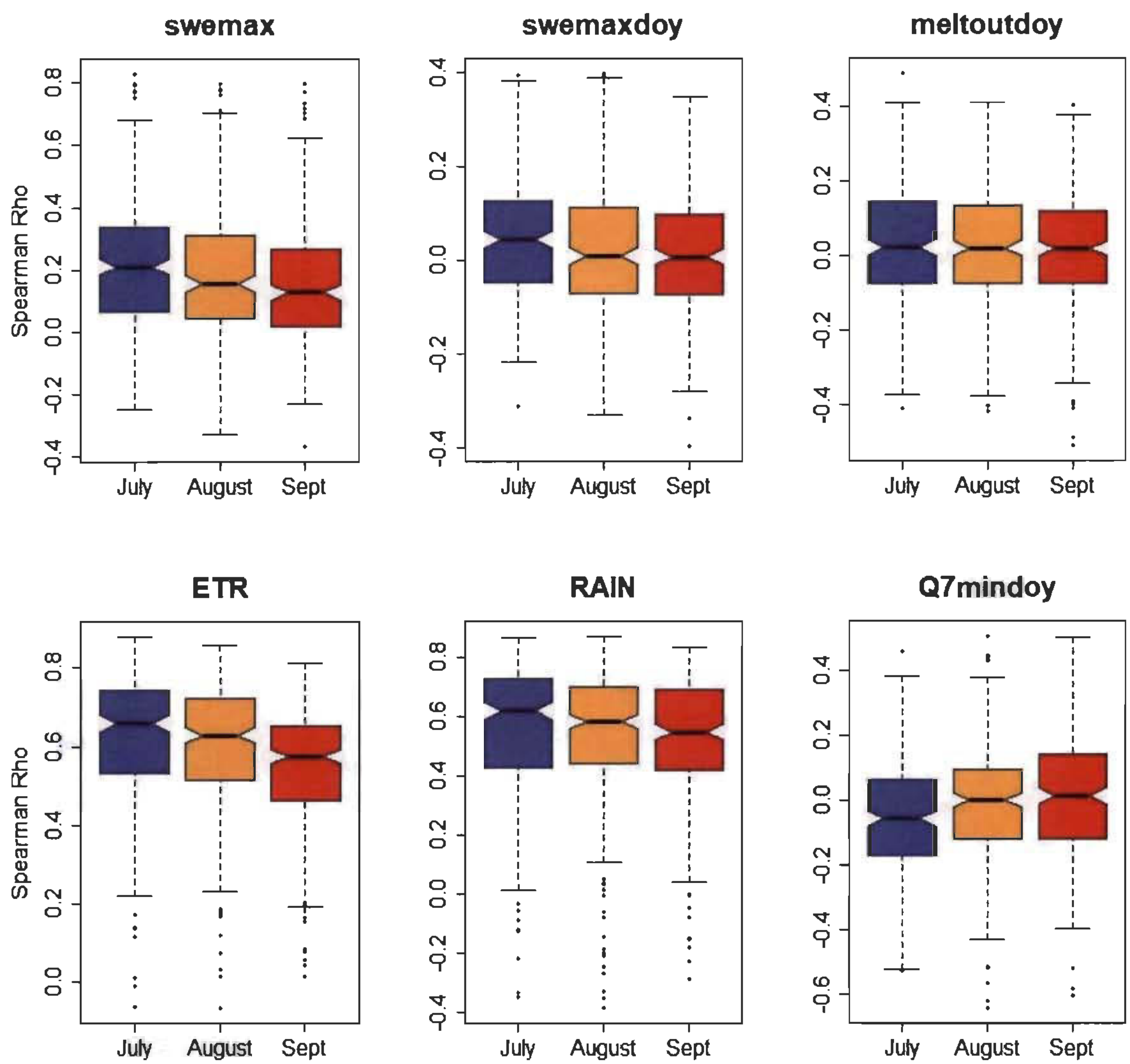


Fig. 2 Boxplots showing the inter basin variability of the Spearman correlation between the six explicative variables and low flow magnitude Q7min during July (blue), August (orange) and September (red).

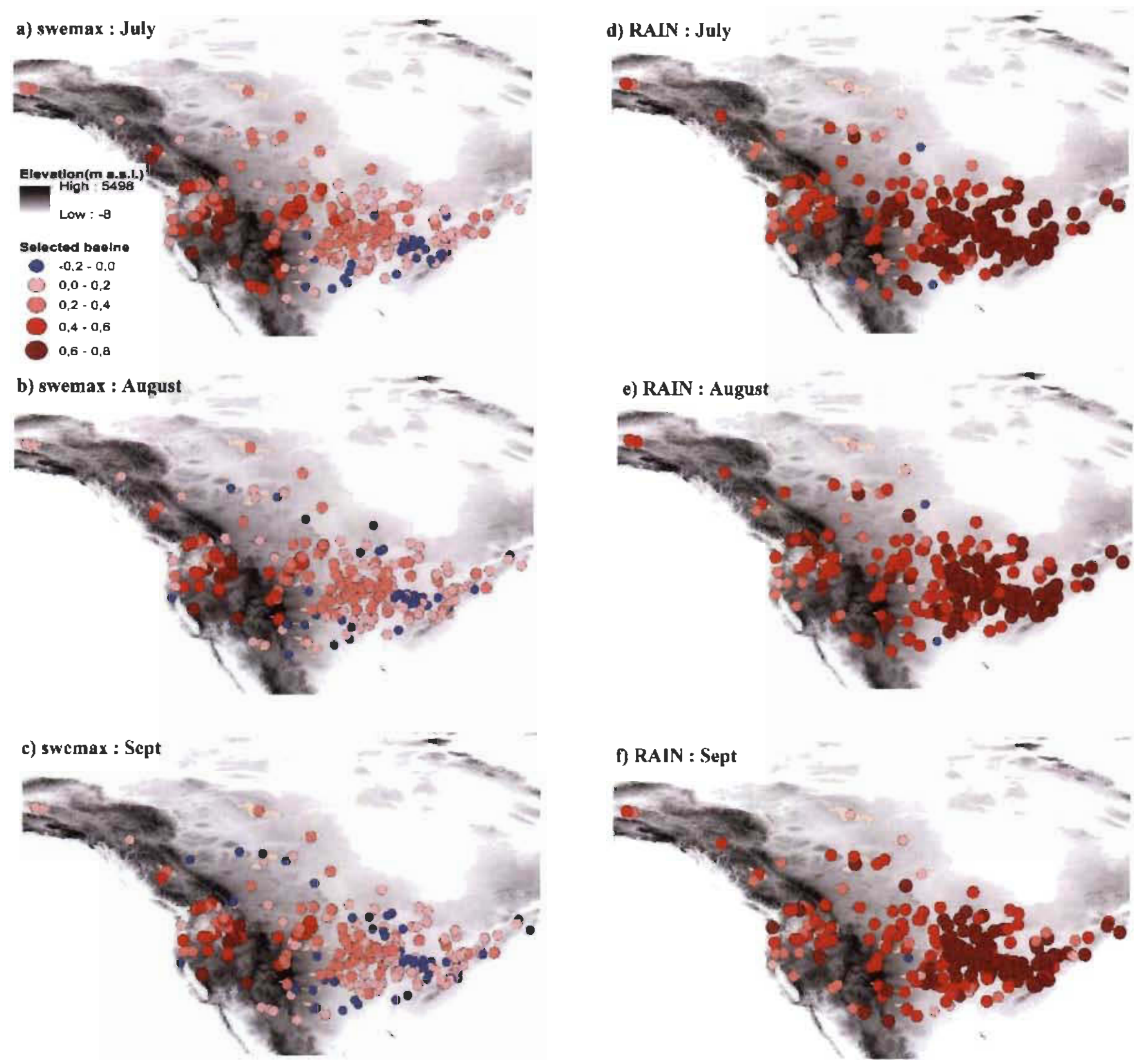


Fig. 3 Maps of correlations between summer low flow and swemax (left: a-c) and summer rainfall (right: e-f) for July, August and September. Significant correlation ($p < 0.05$); Dt size indicate the strength of the correlation. Blue colors indicate negative correlations and red colors positive correlations.

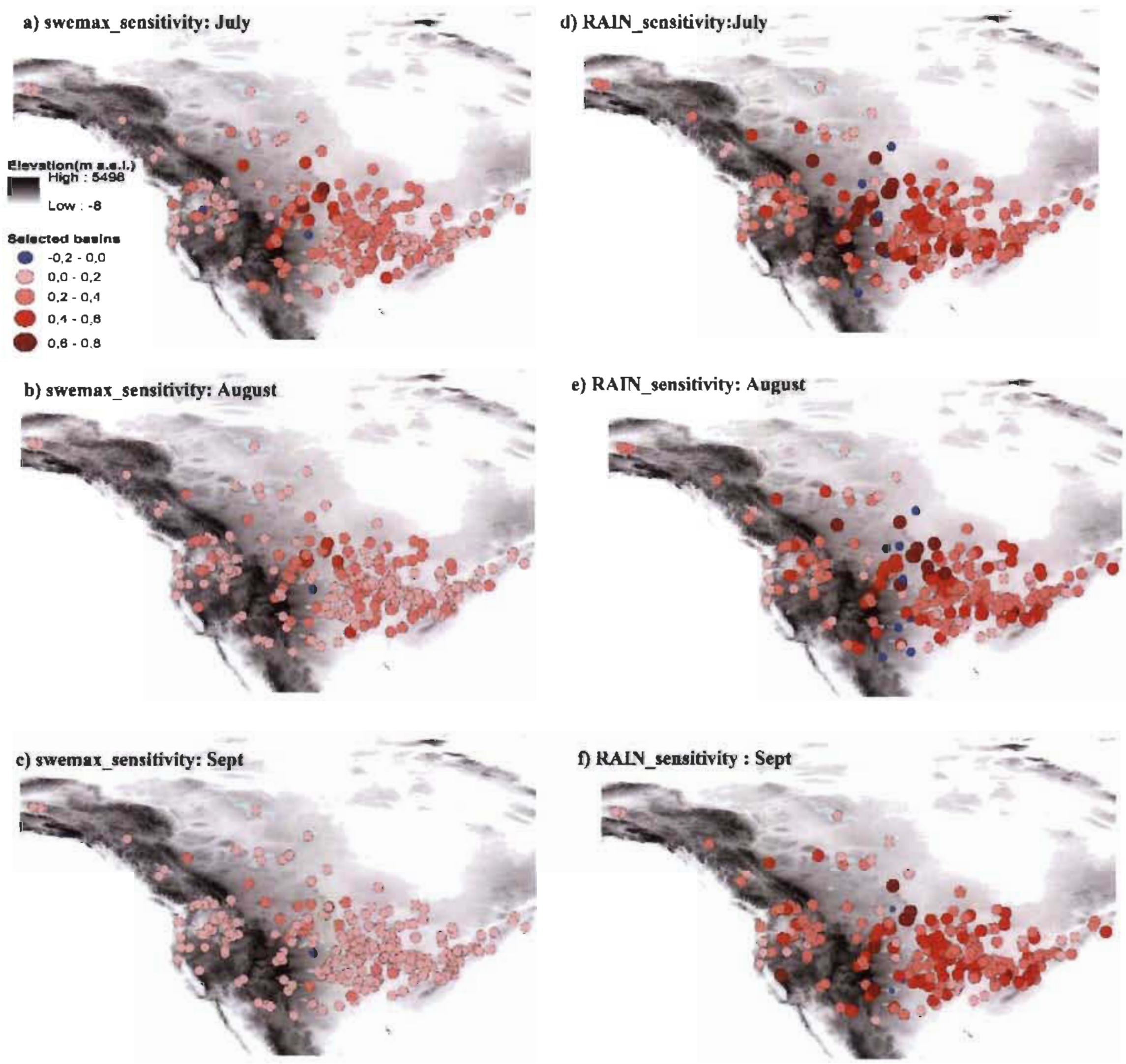


Fig. 4 Maps of the sensitivity of Q7min on swemax (left: a-c) and summer rainfall (right: e-f) for July, August and September. Significant correlation ($p < 0.05$); Dot size indicate the strength of the correlation. Blue colors indicate negative correlations and red colors positive correlations.

References

Abatzoglou, J. T., Dobrowski, S. Z., Parks, S. A., & Hegewisch, K. C. (2018). TerraClimate, a high-resolution global dataset of monthly climate and climatic water balance from 1958–2015. Scientific data, 5, 170191.

Anctil, F., Larouche, W., & Hoang, V. D. (2000). Regional analysis of low water levels. 7-days in the province of Quebec]. Water Quality Research Journal of Canada, 35(1), 125-146.

Assani, A., Gravel, É., Buffin-Bélanger, T., & Roy, A. (2005). Impacts des barrages sur les débits annuels minimums en fonction des régimes hydrologiques artificialisés au Québec (Canada). Revue des sciences de l'eau / Journal of Water Science, 18(1), 103-127. doi:https://doi.org/10.7202/705552ar

Barnett, T. P., Adam, J. C., & Lettenmaier, D. P. (2005). Potential impacts of a warming climate on water availability in snow-dominated regions. Nature, 438(7066), 303-309. doi:10.1038/nature04141

Barnhart, T. B., Molotch, N. P., Livneh, B., Harpold, A. A., Knowles, J. F., & Schneider, D. (2016). Snowmelt rate dictates streamflow. Geophysical Research Letters, 43(15), 8006-8016.

Beaulieu, M., Schreier, H., & Jost, G. (2012). A shifting hydrological regime: a field investigation of snowmelt runoff processes and their connection to summer base flow, Sunshine Coast, British Columbia. Hydrological Processes, 26(17), 2672-2682.

Berghuijs, W. R., Woods, R. A., & Hrachowitz, M. (2014). A precipitation shift from snow towards rain leads to a decrease in streamflow. Nature Climate Change, 4(7), 583-586. doi:10.1038/nclimate2246

Birsan, M.-V., Molnar, P., Burlando, P., & Pfaundler, M. (2005). Streamflow trends in Switzerland. Journal of Hydrology, 314(1-4), 312-329.

Bosilovich, M., Lucchesi, R., & Suarez, M. (2015). MERRA-2: File specification.

Burn, D. H., Buttle, J. M., Caissie, D., MacCulloch, G., Spence, C., & Stahl, K. (2008). The processes, patterns and impacts of low flows across Canada. Canadian Water Resources Journal, 33(2), 107-124.

Caissie, D., Jolicoeur, S., Bouchard, M., & Poncet, E. (2002). Comparison of streamflow between pre and post timber harvesting in Catamaran Brook (Canada). Journal of Hydrology, 258(1-4), 232-248.

Danielson, J. J., & Gesch, D. B. (2011). Global multi-resolution terrain elevation data 2010 (GMTED2010): US Department of the Interior, US Geological Survey.

Dierauer, J., Whitfield, P., & Allen, D. (2018). Climate Controls on Runoff and Low Flows in Mountain Catchments of Western North America. Water Resources Research, 54. doi:10.1029/2018WR023087.

Earman, S., Campbell, A. R., Phillips, F. M., & Newman, B. D. (2006). Isotopic exchange between snow and atmospheric water vapor: Estimation of the snowmelt component of groundwater recharge in the southwestern United States. Journal of Geophysical Research: Atmospheres, 111(D9).

Freeze, R. A. (1974). Streamflow generation. Reviews of Geophysics, 12(4), 627-647.

Gelaro, R., McCarty, W., J. Suárez, M., Todling, R., Molod, A., Takacs, L., . . . Zhao, B. (2017). The Modern-Era Retrospective Analysis for Research and Applications, Version 2 (MERRA-2) (Vol. 30).

Gelman, A., & Hill, J. (2006). Data Analysis Using Regression and Multilevel/Hierarchical Models (Analytical Methods for Social Research). Cambridge: Cambridge University Press. doi:10.1017/CBO9780511790942.

Godsey, S. E., Kirchner, J. W., & Tague, C. L. (2014). Effects of changes in winter snowpacks on summer low flows: case studies in the Sierra Nevada, California, USA. Hydrological Processes, 28(19), 5048-5064. doi:10.1002/hyp.9943.

Gorelick, N., Hancher, M., Dixon, M., Ilyushchenko, S., Thau, D., & Moore, R. (2017). Google Earth Engine: Planetary-scale geospatial analysis for everyone. Remote Sensing of Environment, 202, 18-27.

Grabs, W., De Couet, T., & Pauler, J. (1996). Freshwater Fluxes from Continents Into the World Oceans: Based on Data of the Global Runoff Data Base: Weltdatenbank Abflu← Bundesanstalt für Gewässerkunde.

Hodgkins, G. A., Dudley, R. W., & Huntington, T. G. (2005). SUMMER LOW FLOWS IN NEW ENGLAND DURING THE 20TH CENTURY 1. JAWRA Journal of the American Water Resources Association, 41(2), 403-411.

Jasechko, S., Birks, S. J., Gleeson, T., Wada, Y., Fawcett, P. J., Sharp, Z. D., . . . Welker, J. M. (2014). The pronounced seasonality of global groundwater recharge. Water Resources Research, 50(11), 8845-8867.

Jenicek, M., & Ledvinka, O. (2020). Importance of snowmelt contribution to seasonal runoff and summer low flows in Czechia. Hydrology and Earth System Sciences, 24(7), 3475-3491.

Jenicek, M., Seibert, J., Zappa, M., Staudinger, M., & Jonas, T. (2016). Importance of maximum snow accumulation for summer low flows in humid catchments. Hydrology and Earth System Sciences, 20(2), 859-874. doi:10.5194/hess-20-859-2016.

Laghari, A. N., Vanham, D., & Rauch, W. (2012). To what extent does climate change result in a shift in Alpine hydrology? A case study in the Austrian Alps. Hydrological Sciences Journal, 57(1), 103-117. doi:10.1080/02626667.2011.637040.

Lehner, B., Verdin, K., & Jarvis, A. (2008). New global hydrography derived from spaceborne elevation data. Eos, Transactions American Geophysical Union, 89(10), 93-94.

Mankin, J., Viviroli, D., Singh, D., Hoekstra, A., & Diffenbaugh, N. (2015). The potential for snow to supply human water demand in the present and future. Environmental Research Letters, 10, 114016. doi:10.1088/1748-9326/10/11/114016.

Morice, C. P., Kennedy, J. J., Rayner, N. A., & Jones, P. D. (2012). Quantifying uncertainties in global and regional temperature change using an ensemble of observational estimates: The HadCRUT4 data set. Journal of Geophysical Research: Atmospheres, 117(D8).

Mumma, M., Assani, A., Landry, R., Quessy, J., & Mesfioui, M. (2001). Effect of the change from forest to agriculture land use on the spatial variability of summer extreme daily flow characteristics in Southern Quebec (Canada). Journal of Hydrology, 407(1-4), 153-163.

Neary, D. G., Ice, G. G., & Jackson, C. R. (2009). Linkages between forest soils and water quality and quantity. Forest ecology and management, 258(10), 2269-2281.

New, M., Hulme, M., & Jones, P. (2000). Representing twentieth-century space-time climate variability. Part II: Development of 1901–96 monthly grids of terrestrial surface climate. Journal of Climate, 13(13), 2217-2238.

Pfeffer, W. T., Arendt, A. A., Bliss, A., Bolch, T., Cogley, J. G., Gardner, A. S., . . . Kienholz, C. (2014). The Randolph Glacier Inventory: a globally complete inventory of glaciers. Journal of Glaciology, 60(221), 537-552.

Pike, R., & Scherer, R. (2003). Overview of the potential effects of forest management on low flows in snowmelt-dominated hydrologic regimes. Journal of Ecosystems and Management, 3(1).

Pomeroy, J., Fang, X., & Ellis, C. (2012). Sensitivity of snowmelt hydrology in Marmot Creek, Alberta, to forest cover disturbance. Hydrological Processes, 26(12), 1891-1904.

Pomeroy, J. W., Spence, C., & Whitfield, P. H. (2013). Putting Prediction in Ungauged Basins Into Practice: Canadian Water Resources Association.

Price, K. (2011). Effects of watershed topography, soils, land use, and climate on baseflow hydrology in humid regions: A review. Progress in physical geography, 35(4), 465-492.

Quilbé, R., Rousseau, A. N., Moquet, J.-S., Savary, S., Ricard, S., & Garbouj, M. S. (2008). Hydrological responses of a watershed to historical land use evolution and future land use scenarios under climate change conditions.

Roche, M. F. (1986). Dictionnaire français d'hydrologie de surface avec équivalents en anglais-espagnol-allemand: Masson.

Sakia, R. M. (1992). The Box-Cox transformation technique: a review. Journal of the Royal Statistical Society: Series D (The Statistician), 41(2), 169-178.

Smakhtin, V. U. (2001). Low flow hydrology: a review. Journal of Hydrology, 240(3), 147-186. doi:https://doi.org/10.1016/S0022-1694(00)00340-1

Staudinger, M., Stoelzle, M., Seeger, S., Seibert, J., Weiler, M., & Stahl, K. (2017). Catchment water storage variation with elevation. Hydrological Processes, 31. doi:10.1002/hyp.11158

Stewart, I. T., Cayan, D. R., & Dettinger, M. D. (2004). Changes in snowmelt runoff timing in western North America under abusiness as usual'climate change scenario. Climatic Change, 62(1-3), 217-232.

Sulla-Menashe, D., & Friedl, M. A. (2018). User guide to collection 6 MODIS land cover (MCD12Q1 and MCD12C1) product. USGS: Reston, VA, USA, 1-18.

Sun, N., Wigmosta, M., Zhou, T., Lundquist, J., Dickerson-Lange, S., & Cristea, N. (2018). Evaluating the functionality and streamflow impacts of explicitly modelling forest–snow interactions and canopy gaps in a distributed hydrologic model. Hydrological Processes, 32(13), 2128-2140.

Sylvain, J.-M., Assani, A., Landry, R., Quessy, J.-F., & Kinnard, C. (2015). Comparison of the spatio-temporal variability of annual minimum daily extreme flow characteristics as a function of land use and dam management mode in Quebec, Canada. Water, 7(3), 1232-1245.

Tague, C., & Grant, G. E. (2009). Groundwater dynamics mediate low-flow response to global warming in snow-dominated alpine regions. Water Resources Research, 45(7).

Tallaksen, L. (1995). A review of baseflow recession analysis. Journal of Hydrology, 165(1-4), 349-370.

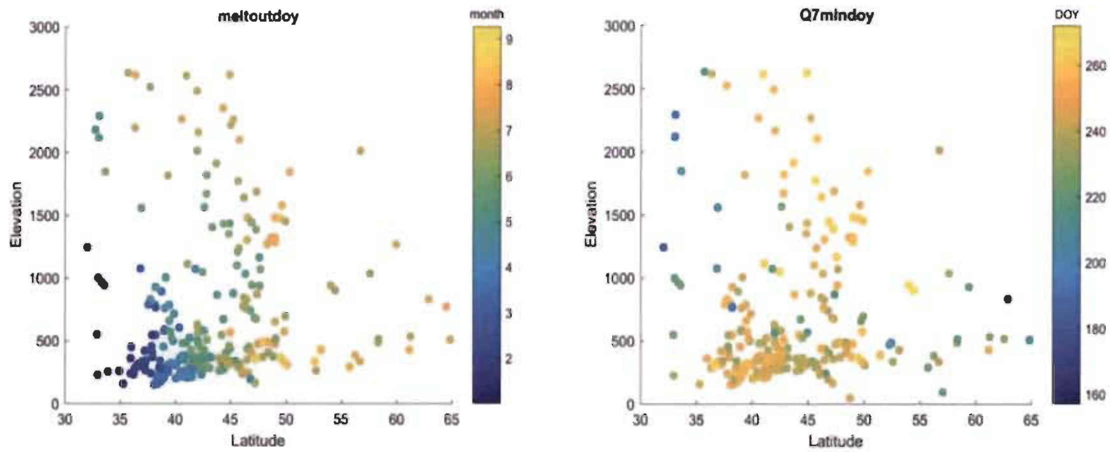
Van Loon, A., & Laaha, G. (2015). Hydrological drought severity explained by climate and catchment characteristics. Journal of Hydrology, 526, 3-14.

Yuan, W., Liu, S., Yu, G., Bonnefond, J.-M., Chen, J., Davis, K., . . . Rossi, F. (2010). Global estimates of evapotranspiration and gross primary production based on MODIS and global meteorology data. Remote Sensing of Environment, 114(7), 1416-1431.

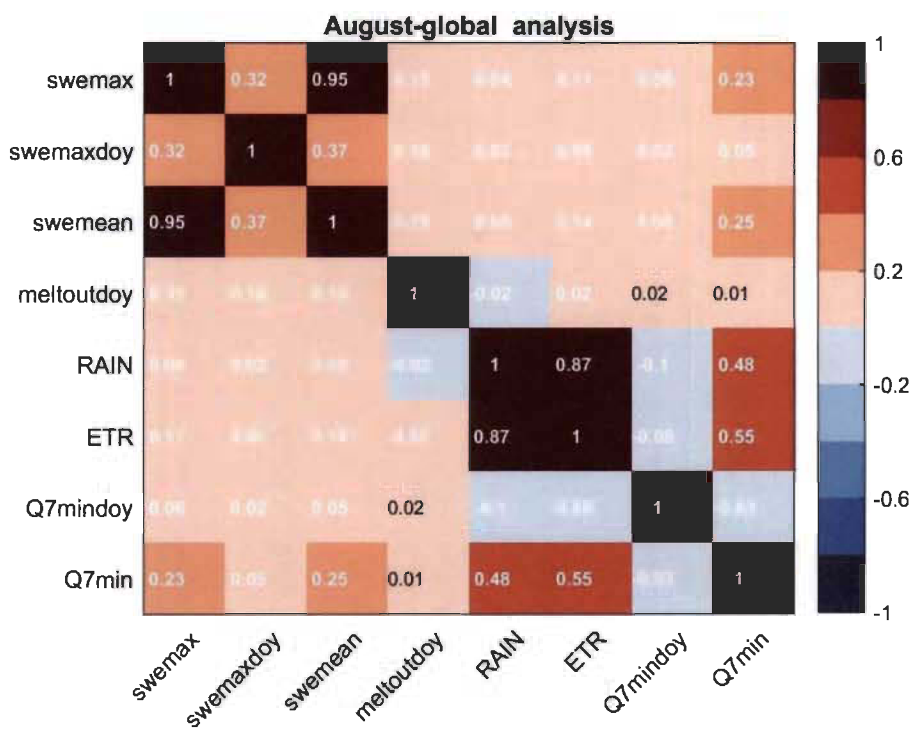
Yuan, Z., & Chen, H. Y. (2010). Fine root biomass, production, turnover rates, and nutrient contents in boreal forest ecosystems in relation to species, climate, fertility, and stand age: literature review and meta-analyses. Critical Reviews in Plant Sciences, 29(4), 204-221.

Zuur, A., Ieno, E. N., Walker, N., Saveliev, A. A., & Smith, G. M. (2009). Mixed effects models and extensions in ecology with R: Springer Science & Business Media.

Supplementary material



Supplementary Fig. 1 Inter-annual variability of snow free timing (month) and Inter-annual variability of low flow *Q7min* timing observed in the 260 basins in function of elevation and latitude.



Supplementary Fig. 2 Heatmap showing Spearman correlation coefficient for all antecedent factors and summer low flow Q7min for the 260 basins in August. Significant correlation ($p < 0.05$) are highlighted in white; blue colors indicate negative correlations and brown colors positive correlations.

CHAPITRE IV

CONCLUSION GÉNÉRALE

L'objectif principal de ce projet était d'examiner l'impact des variations interannuelles de l'accumulation de neige en hiver sur les débits d'étiage en été dans les régions froides de l'hémisphère nord et de comparer la contribution respective des variables liées à l'accumulation hivernale de neige et de son taux de fonte au printemps à la variabilité interannuelle du débit d'étiage, par rapport à celles de l'évapotranspiration et de la pluie estivale. Pour évaluer les relations entre les prédicteurs et les variables de réponse, nous avons utilisé le coefficient de corrélation de rang de Spearman et la régression multiniveaux. Pour ceci la plupart des prédicteurs et des variables de réponse ont été préalablement standardisés et normalisés, ce qui a permis de comparer les différents bassins versants entre eux.

La contribution de la première partie de cette étude (Chapitre II) est une première analyse réalisée sur un échantillon de 12 bassins-versants des affluents du fleuve Saint-Laurent au sud Québec avec un régime naturel et une variabilité significative dans les précipitations hivernales et estivales. Les principaux résultats montrent que : la corrélation bivariée globale entre les différents facteurs météorologiques hivernaux et estivaux sur les débits observés a montré que globalement, l'équivalent en eau de neige maximal accumulé pendant l'hiver est un bon prédicteur de Q7min en été, tandis que la variabilité interannuelle du Q7min est moins affecté par la date d'occurrence de swemax, la date de disparition de la neige ainsi que le taux de fonte. Par conséquent, en hiver sec, nous nous attendons à une diminution (magnitude) des faibles débits en été. Cependant, l'analyse de corrélation par bassin a révélé des hétérogénéités significatives entre ces bassins. Dans la plupart des bassins étudiés l'effet de swemax sur Q7min est plus important vers la fin de l'été (août/septembre). D'où, le débit d'étiage retardé pendant l'été est alimenté par la recharge prédominante des aquifères par la fonte des neiges printanière suivie de l'écoulement progressif des aquifères vers les cours d'eau en été.

Cependant, seulement dans les deux bassins au climat maritime, une accumulation importante de neige en hiver était lié à une faible baisse du volume des débits d'étiages en été. Comme prévu, globalement, la quantité de pluie accumulée entre la date de disparition de la neige et la date du débit minimum estival est le meilleur prédicteur de Q7min, les étés plus humides mènent à des débits plus faibles en été. Les pertes d'évapotranspiration sont associées à une diminution des faibles débits, mais la corrélation est plus faible que celle pour la pluie pendant l'été et l'accumulation de neige en hiver. De plus, la corrélation bivariée a montré aussi que Q7min n'est pas corrélé avec sa date d'occurrence, de sorte que des débits plus faibles n'ont pas tendance à se produire plus tard dans l'année. Le début des chutes de pluie automnales dans le sud du Québec est le contrôle le plus probable du moment des minimums de faible débit en été dans le sud du Québec. Cependant, l'analyse multivariée a montré que la date d'occurrence plus tardive est liée à un volume plus faible de Q7min. D'où il est important d'utiliser une approche multivariée afin d'éviter l'effet croisé des prédicteurs.

L'approche de régression hiérarchique utilisée montrent que les variations interannuelles des précipitations estivales et des pertes par évapotranspiration sont le principal contrôle de la variabilité estivale, et que les conditions hivernales antérieures représentent toujours un rôle important mais secondaire de variabilité interannuelle dans le sud du Québec. Les résultats montrent aussi que, la date d'occurrence plus tardive est liée à une récession du volume de débit pendant l'été. En général, les résultats de corrélation (globale et par bassin) ne donnent pas toujours les mêmes relations que l'analyse multivariée. Les prédicteurs (swemax et RAIN) semblent avoir des relations cohérentes avec Q7min, alors que d'autres variables ont des effets différents lorsque l'influence des autres prédicteurs est prise en compte dans le modèle mixte. Parmi les descripteurs du bassin inclus pour expliquer cette variabilité entre les bassins, on a constaté que seulement le couvert forestier avait une influence positive importante sur Q7min. Par conséquent, dans les bassins boisés le débit d'étiage est plus élevé, surtout vers la fin de l'été.

La deuxième partie de cette étude (Chapitre III) est une analyse globale inspirée des résultats et méthodes développées dans la première partie sur un large échantillon de bassins dans l'Amérique du nord où le ruissèlement est dominé par la fonte de neige pendant le printemps.

Les principaux résultats montrent que globalement, l'accumulation de la neige pendant l'hiver a considérablement affecté le volume de faible débit en été, mais cet effet est différent d'un bassin à un autre et d'un mois à un autre. L'effet du stockage nival sur le débit d'étiage est plus important dans le mois de juillet et diminue vers la fin de l'été.

La sensibilité de Q7min sur swemax a été affectée par plusieurs descripteurs de bassins. La contribution de la fonte des neiges aux débits détiage en été est plus importante dans les bassins montagneux que dans les Prairies vers l'est. De plus, la latitude a un effet positif sur la sensibilité de la variabilité interannuelle de Q7min sur swemax. Ainsi, les basins dans le sud-est du continent nord-américain ont montré un effet négligeable de swemax sur Q7min. Dans les bassins boisés le débit d'étiage est moins sensible à la variabilité interannuelle de stockage nivale, possiblement en raison de l'interception de la neige par la canopée qui réduit l'accumulation de neige au sol. Généralement, les pluies estivales ont l'effet le plus important sur le débit d'étiage estival, avec un patron spatial inversé à celui de l'accumulation maximale de la neige. La contribution de pluie au volume de Q7min est plus importante dans les bassins à climat continental humide, c.-à-d. dans le centre-est de l'Amérique du Nord et plus faible dans la région montagneuse et semi-aride de l'Ouest. Ainsi, il y a encore une partie de la variabilité Q7min expliqué par les descripteurs des bassins. En effet, dans les bassins boisés la magnitude Q7min plus élevée que les bassins agricoles. Malgré la haute résolution de l'ensembles de bases des données globales utilisé dans cette étude certaines incertitudes existent dans l'estimation des données météorologiques (swemax, précipitations, évapoyranspiration) liées à la mesure hétérogène, à l'échantillonnage non corrélées et l'utilisation un modèle de bilan hydrique simple basé uniquement sur une combinaison des variables climatiques essentielles sans tenant compte de l'hétérogénéité des types de végétation, ce qui abaissent la capacité prédictive du modèle statistique utilisé et rendent les faibles débits prévus potentiellement non réaliste. Par conséquent, pour d'améliorer notre compréhension des mécanismes de faibles débits dans les pays nordique, il est recommandé d'utiliser de modèles empiriques pour les précipitations et la température avec les données régionales de réanalyse, la modélisation de l'humidité du sol, ainsi que les facteurs topographiques et la couverture de la végétales

RÉFÉRENCES BIBLIOGRAPHIQUES

Assani, A., Gravel, É., Buffin-Bélanger, T., & Roy, A. (2005). Impacts des barrages sur les débits annuels minimums en fonction des régimes hydrologiques artificialisés au Québec (Canada). Revue des sciences de l'eau / Journal of Water Science, 18(1), 103-127. doi:https://doi.org/10.7202/705552ar

Assani, A. A., Chalifour, A., Légaré, G., Manouane, C.-S., & Leroux, D. (2011). Temporal Regionalization of 7-Day Low Flows in the St. Lawrence Watershed in Quebec (Canada). Water Resources Management, 25(14), 3559-3574. doi:10.1007/s11269-011-9870-6

Assani, A. A., Charron, S., Matteau, M., Mesfioui, M., & Quessy, J.-F. (2010). Temporal variability modes of floods for catchments in the St. Lawrence watershed (Quebec, Canada). Journal of Hydrology, 385(1), 292-299. doi:https://doi.org/10.1016/j.jhydrol.2010.02.031

Aygün, O., Kinnard, C., & Campeau, S. (2020). Impacts of climate change on the hydrology of northern midlatitude cold regions. Progress in Physical Geography: Earth and Environment, 44(3), 338-375.

Barnett, T. P., Adam, J. C., & Lettenmaier, D. P. (2005). Potential impacts of a warming climate on water availability in snow-dominated regions. Nature, 438(7066), 303-309. doi:10.1038/nature04141

Berghuijs, W. R., Woods, R. A., & Hrachowitz, M. (2014). A precipitation shift from snow towards rain leads to a decrease in streamflow. Nature Climate Change, 4(7), 583-586. doi:10.1038/nclimate2246

Bradford, M. J., & Heinonen, J. S. (2008). Low flows, instream flow needs and fish ecology in small streams. Canadian Water Resources Journal, 33(2), 165-180.

Feng, S., & Hu, Q. (2007). Changes in winter snowfall/precipitation ratio in the contiguous United States. Journal of Geophysical Research: Atmospheres, 112(D15).

GIEC, R., & Pachauri, A. R. (2014). Changements Climatiques 2014: Rapport de Synthèse. In: GIEC.

Godsey, S. E., Kirchner, J. W., & Tague, C. L. (2014). Effects of changes in winter snowpacks on summer low flows: case studies in the Sierra Nevada, California, USA. Hydrological Processes, 28(19), 5048-5064. doi:10.1002/hyp.9943

Jenicek, M., Seibert, J., Zappa, M., Staudinger, M., & Jonas, T. (2016). Importance of maximum snow accumulation for summer low flows in humid catchments. Hydrology and Earth System Sciences, 20(2), 859-874. doi:10.5194/hess-20-859-2016

Smakhtin, V. U. (2001). Low flow hydrology: a review. Journal of Hydrology, 240(3), 147-186. doi:https://doi.org/10.1016/S0022-1694(00)00340-1

Walters, A. W. (2016). The importance of context dependence for understanding the effects of low-flow events on fish. Freshwater Science, 35(1), 216-228.

Zhang, F.-Y., Li, L.-H., Ahmad, S., & Li, X.-M. (2014). Using path analysis to identify the influence of climatic factors on spring peak flow dominated by snowmelt in an alpine watershed. Journal of Mountain Science, 11(4), 990-1000. doi:10.1007/s11629-013-2789-z