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Characterization of contrasting flow and thermal regimes in two adjacent subarctic alpine headwaters in northwest Canada

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Data Availability Statement

Plots depicting the data that support the findings of this study are available in the supplementary material of this article, while numerical data are available from the corresponding author upon reasonable request.

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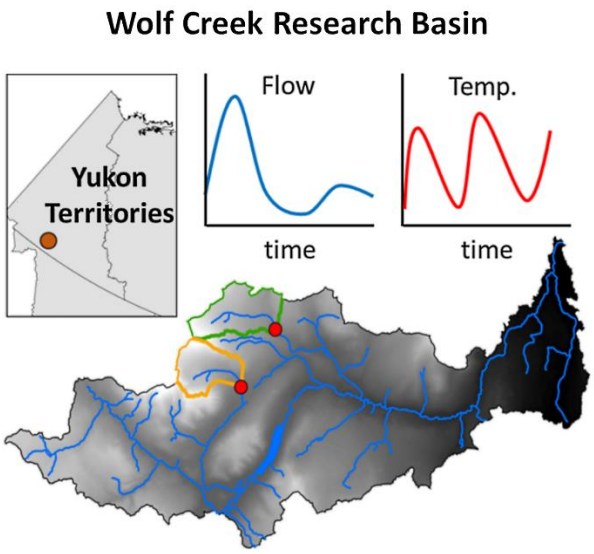
Key words: flow regime; stream temperature; heat budget; energy balance; subarctic alpine headwaters; regression analysis; permafrost; climate change

Abstract

Alpine headwaters in subarctic regions are particularly sensitive to climate change, yet there is little information on stream thermal regimes in these areas and how they might respond to global warming. In this paper, we characterize and compare the hydrological and thermal regimes of two subarctic headwater alpine streams within an empirical framework. The streams investigated are located within two adjacent catchments with similar geology, size, elevation and landscape, Granger Creek (GC) and Buckbrush Creek (BB), which are part of the Wolf Creek Research Basin in the Yukon Territory, Canada. Hydrometeorological and high-resolution stream temperature data were collected throughout summer 2016. Both sites exhibited a flow regime typical of cold alpine headwater catchments influenced by frozen ground and permafrost. Comparatively, GC was characterized by a flashier response with more extreme flows, than BB. In both sites, stream temperature was highly variable and very responsive to short-term changes in climatic conditions. On average, stream temperature in BB was slightly higher than in GC (respectively 5.8 °C and 5.7 °C), but less variable (average difference between 75^{th} - 25^{th} quantiles of 1.6 °C and 2.0 °C). Regression analysis between mean daily air and stream

temperature suggested that a greater relative (to stream flow) groundwater contribution in BB could more effectively buffer atmospheric fluctuations. Heat fluxes were derived and utilized to assess their relative contribution to the energy balance. Overall, non-advective fluxes followed a daily pattern highly correlated to short-wave radiation. Generally, solar radiation and latent heat were respectively the most important heat source and sink, while air-water interface processes were major factors driving nighttime stream temperature fluctuations.



Graphical abstract

1 INTRODUCTION

Stream temperature is a key factor regulating chemical, physical and biological processes within rivers (Webb, Hannah, Moore, Brown & Nobilis, 2008), such as the distribution and growth rate of aquatic organisms (Marine & Cech, 2004), concentration of dissolved oxygen and sediments (Ficklin, Stewart & Maurer, 2013), and the release of nutrients (Mcdowell, Elkin & Kleinman, 2017). Consequently, riverine ecosystems are very sensitive to alterations of diel and seasonal stream thermal cycles and spatial temperature patterns (Fullerton *et al*., 2018; Kurylyk *et al.*, 2015; Steel *et al*., 2017). These alterations can be the result of natural changes, anthropogenic

perturbations, or their combinations (Caissie, 2006). The main anthropogenic perturbations include deforestation (e.g. Bourque & Pomeroy, 2001), flow modifications (e.g. construction of water reservoirs and canalization, Ferrazzi & Botter, 2019; Gaay & Blokland, 1970; Liu, Yang, Ye & Berezovskaya, 2005), urban water inputs and withdrawals (e.g. Kinouchi, Yagi & Miyamoto, 2007), and climate change (e.g. Ficklin *et al*., 2013; Isaak, Wollrab, Horan & Chandler, 2012).

Presently, climate change represents the most concerning threat for river thermal environments at a global scale, with considerable research directly linking long-term increases in stream temperatures to global warming (e.g. Arora, Tockner & Venohr, 2016; Hari, Livingstone, Siber, Burkhardt-Holm & Guttinger, 2006; Isaak *et al.*, 2012; Mayer, 2012). Stream temperature is the result of complex interactions between heat fluxes at the air-water interface (radiative and turbulent fluxes) (Caissie, 2006; Hebert, Caissie, Satish & El-jabi, 2011; Leach & Moore, 2010), streambed conduction (Caissie & Luce, 2017), and advective processes occurring along the channel with streamflow, across the streambed interface due to groundwater-surface water interactions (Briggs *et al.*, 2018; Caissie & Luce, 2017; Kurylyk, Moore & Macquarrie, 2016), or transversely due to runoff, bank seeps or springs, or tributary inflows (Harrington, Hayashi & Kurylyk, 2017; Leach & Moore, 2014). Climate change will alter the thermal cycle in two distinct ways: directly, by altering the surface energy fluxes, which normally dominate a stream's energy budget; and indirectly, by changing hydrological conditions and thus advective fluxes and thermal inertia (Pekarova, Halmova, Miklanek, Onderka, Pekar & Skoda., 2008; Poole & Berman, 2001; Vliet, Ludwig, Zwolsman, Weedon & Kabat, 2011).

The magnitude and temporal patterns of these thermal and hydrologic changes are not uniformly distributed, but are strongly related to nested controls acting at different spatial and temporal

scales, including climate (first order), basin characteristics (second order) and reach-specific controls (third order) (Hannah & Garner, 2015; Webb *et al.*, 2008). Understanding how these conditions vary among watersheds may help to explain why some rivers are warming more slowly than others (Isaak *et al.*, 2016; Luce, Staab, Kramer, Wenger, Isaak & McConnell, 2014), and why some river networks or catchments that are historically more thermally resilient are expected to provide future climate refugia and preserve cold-water species (Isaak, Young, Nagel, Horan & Groce, 2015). Therefore, if we wish to improve our capability to predict future patterns and design efficient mitigation and conservation strategies, understanding the hydrological and thermal cycles, and heat fluxes regulating stream temperature in different climatic regions and environmental conditions is fundamental (Macdonald, Boon, Byrne & Silins, 2014).

Headwaters are generally characterized by high thermal heterogeneity, low thermal inertia, and variable sources of water inputs (Brown, Hannah & Milner, 2005; Brown & Hannah, 2008; Mayer, 2012; Harrington *et al.*, 2017; Docherty, Abermann, Lund & Hannah, 2019). Of major importance in these streams is the discharge of groundwater (Mayer, 2012), which by acting as a heat source and sink in cold and warm periods respectively, reduces their thermal sensitivity to change in climatic conditions (e.g. air temperature and shortwave radiation) (Evans & Petts, 1997; Poole *et al*., 2008). This results in streams with large groundwater contribution (with respect to longitudinal flow) having reduced temperature variability at multiple time scales (e.g. diel, seasonal) (Kelleher, Wagener, Gooseff, Mcglynn, Mcguire & Marshall., 2012) , but there is some debate on whether groundwater-dominated systems will also exhibit great resilience to multi-decadal climate warming (Kaandorp, Doornenbal, Kooi, Peter Broers & de Louw, 2019; Kurylyk, MacQuarrie & Voss, 2014). Additionally, the focused discharge of groundwater creates

thermal anomalies that provide temporary thermal refugia in both the summer and winter (Power, Brown & Imhof, 1999).

Headwater streams in subarctic regions (dominated by permafrost) are particularly susceptible to changes in their thermal regimes, as future climate projections agree that the greatest warming will continue to be observed at both high latitudes and altitudes (Debeer, Wheater, Carey & Chun, 2016; Pepin *et al*., 2015). Nevertheless, the bulk of the research has focused on valley bottom rivers (e.g. Broadmeadow, Jones, Langford, Shaw & Nisbet, 2011; Ficklin *et al*., 2013; Kinouchi *et al*., 2007) and alpine headwater catchments (e.g. Khamis, Brown, Milner & Hannah, 2015; Leach & Moore, 2017; Webb *et al*., 2008) in temperate regions. Conversely, studies in arctic and subarctic regions are rather rare (e.g. King, Neilson, Overbeck & Kane, 2016; King & Neilson, 2019; Docherty *et al*., 2019).

Another important impact of global warming in cold regions is related to the presence of permafrost and snow accumulated during the cold season. Permafrost is a key control of many hydrological processes such as water storage and connectivity (interaction between shallow and deep subsurface water). The risk of increased permafrost degradation and decreased seasonal ground frost under future climate change (DeBeer *et al*., 2016; Okkonen, Jyrkama & Kløve, 2010) have contributed to an increase of studies conducted on rivers and streams in these regions (e.g. Walvoord and Kurylyk, 2016). Decreases in seasonal or perennial ground ice could potentially lead to more active groundwater systems and deeper routing of subsurface water, causing these systems to gradually exhibit more of the characteristics of groundwater-dominated rivers (Quinton, Hayashi & Chasmer, 2009; Walvoord, Voss & Wellman, 2012).

In general, in alpine catchments with snowmelt as the primary source feeding baseflow, an increase in air temperature will alter the timing and duration of snowmelt, increasing the

possibility of extremely low flows (Carey, Boucher & Duarte, 2013; MacDonald *et al.* 2014; Okkonen *et al.,* 2010), and thus of extreme water temperatures (Van Vliet, Franssen, Yearsley, Ludwig, Haddeland, Lettenmaier & Kabat, 2013). In contrast, when snowmelt can sustain streamflow throughout the warm season, an increase in air temperature might increase streamflow and reduce its thermal sensitivity to climatic conditions (Lisi, Schindler, Cline, Scheuerell & Walsh, 2015).

Hydrological and thermal cycles in cold alpine headwaters in the subarctic region are poorly characterized, but may have critical ecological function under a changing climate (Caissie, 2006; Gooseff, 2010) by providing a 'cold-water shield' for at-risk coldwater species (Isaak *et al.*, 2015). Accordingly, the overall purpose of this study is to identify the distinctive characteristics of the hydrological and thermal regimes in two adjacent alpine headwaters within the Wolf Creek Research Basin, Yukon Territory (Rasouli, Pomeroy, Janowicz, Williams & Carey, 2019). More specifically the objectives of this paper are to (1) characterize and compare the flow regime dynamics; (2) quantify the relative contribution of different heat fluxes to the stream thermal balance through applying an energy budget approach; and (3) use high-resolution stream temperature data to characterize spatiotemporal thermal patterns.

2 STUDY SITES

Wolf Creek Research Basin (WCRB, 60°32'47" N, 135°11'03" W) is located ~20 km southeast of Whitehorse in Yukon Territory, Canada (Fig. 1a). WCRB has a drainage area of ~179 km^2, with elevation ranging between 710 and 2080 m above sea level (a.s.l.). The region is characterized by a subarctic continental climate, with low precipitation and wide air temperature ranges (Rasouli *et al*., 2019). The average (1981-2010) air temperature measured at the Whitehorse International Airport (706 m a.s.l.) is -0.1 °C, ranging from -15.2 °C (January) to

14.3 °C (July), while average annual precipitation is ~260 mm, of which ~40 % falls as snow. However, due to orographic factors, precipitation in the WCRB at high elevation can be 25 – 35 % greater (Pomeroy Hedstrom & Parviainen, 1998).

Granger Creek (GC) and Buckbrush Creek (BB) are two headwater catchments located in the northeast of WCRB with similar catchment area (GC: 7.6 km^2; BB: 6.1 km^2), elevation range (GC: 1355 - 2080 m a.s.l.; BB: 1324 - 2080 m a.s.l.), and main stem flow length (GC: 4.8 km; BB: 3.8 km; Fig. 1a). Each creek has a common west-east flow direction and bed sediment consisting of boulders, cobbles, and gravels, with limited sandy deposits. This study focuses on a 1950 m reach of each stream with similar channel characteristics (Fig. 1b). The upper sections of the reaches are relatively narrow, characterized by slightly meandering channels with little shading from vegetation (Fig. 1c and 1d for GC and BB, respectively). Conversely, the lower portions are wider, with a step-pool morphology and denser bank vegetation (Fig. 1e and 1f for GC and BB, respectively). One and two tributaries are present in the lower parts of BB and GC, respectively (blue arrows, Fig. 1b).

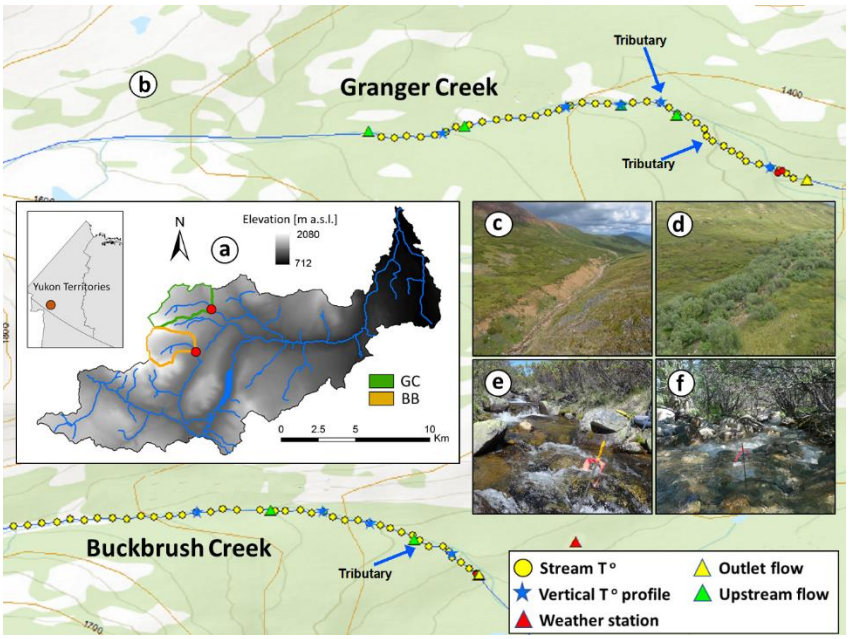


Fig. 1. Wolf Creek Research Basin (WCRB), Granger Creek (GC) and Buckbrush Creek (BB) basins and their locations (a). Study reaches and gauging stations (b). GC upstream (c) and downstream (e) segments. BB upstream (d) and downstream (f) segments.

The geology of these headwater catchments is predominantly sedimentary, and includes limestone, siltstone, sandstone, and conglomerate, overlaid by till (Mougeot and Smith, 1994). Permafrost underlies much of the north-facing aspects, while seasonal frost is dominant in the south-facing slopes and at lower elevations (Carey & Quinton 2005; Lewkowicz & Ednie, 2004). Groundwater discharge has been identified as the main source feeding creeks in the alpine zone using both geophysics (Seguin, Stein, Nilo & Jalbert, 1999) and tracer data (Carey *et al*., 2013). Vegetation varies with elevation, with tall shrubs (Salix) dominating in valley bottoms that transitions to shorter shrubs (Salix and Betula) as elevation increases, with tundra foliage and bare rock at the highest elevations.

3 METHODOLOGY

3.1 Field data

The study period spanned the summer field season from 22 June (after snowmelt occurred) to 15 September, 2016. Water temperatures were recorded at a 15-minute frequency using iButtons (model iBWetLand DS1922L, precision of ±0.5 °C, resolution of 0.0625 °C). The loggers were partially shielded in durable silicone and calibrated to 0 ^{o}C in a continuously stirred ice bath. In total, 40 calibrated iButtons were used in each stream, with one installed centre-channel on the bed every 50 m along the length of each 1950 m study reach (yellow points, Fig. 1b). Vertically spaced streambed temperature time-series were recorded at a 30-minute frequency at depths of 0.05 m and 0.15 m below the sediment-water interface by mounting iButtons in survey stakes and driving them into the bed at various locations in each stream (5 in GC and 4 in BB respectively; blue stars, Fig. 1b) to investigate streambed heat fluxes.

Hydrometric stations were established along the length of each creek (5 in GC, 3 in BB), instrumented with Solinst LTC Levelogger Edge pressure transducers to measure stage at 15-minute frequencies (green and yellow triangles, Fig. 1b). Under various flow conditions, water depth and velocity were sampled and utilized to develop stage-discharge relationships for each station according to the velocity-area method (Le Coz, Camenen, Peyrard & Dramais, 2012). Velocity measurements were taken using a SonTek FlowTracker handheld Acoustic Doppler Velocimeter (accuracy ± 1%) at 0.6 of water depths (0.6-depth method is widely accepted as representative of mean column velocity (Genç, Ardıçlıoğlu & Ağıralioğlu, 2015)). For this analysis, we utilized only flow time series estimated at the stream outlets where more accurate rating curves have been built.

Leaf Area Index (LAI) was measured at each sensor location in July using a LI-COR Plant Canopy Analyzer (LAI – 2200C), and upward facing photos were taken with a GoPro Hero3 camera to estimate shading factors (SF). A Matlab code was then developed and applied to these photos to estimate the percentage of the frames covered by canopy. Shading factor was considered to vary spatially but be constant in time.

Meteorological data were collected at a long-term weather station located near BB (red triangle, Fig. 1b). Along with other parameters, the station measured incoming and outgoing shortwave and longwave radiation (Kipp & Zonen, CNR4), relative humidity, air temperature (Campbell Scientific, HMP45C-L), and wind speed (RM Young) at 30-minute intervals. Evapotranspiration (ET) was measured directly using a LI-7500A (Li-Cor) open path infra-red gas analyzer and a three-axis sonic anemometer (R3-50, Windmaster, Gill Instruments) using standard flux protocols (Baldocchi, Hincks & Meyers, 1988). Precipitation was recorded at 30-minute intervals using a tipping-bucket rain gauge (CS700, Campbell Scientific) located nearby the Buckbrush weather station. Some of the main observations used in this study are shown in the supporting material, (Fig. S1).

3.2 Data analysis

3.2.1 Flow characterization

For both streams, flow, rainfall and ET measurements were aggregated at hourly and daily time scales. To allow for comparisons between sites and cross-sections, we considered specific discharge (discharge normalized to basin area, q [mm d^{-1}]). Hourly flows were utilized to construct time series and frequency distributions, while daily discharge, precipitation, and ET were used to derive cumulative frequencies and to estimate water deficits. Overall, these provided a robust base from which to analyze flow patterns and form inter-site comparisons.

3.2.2 Energy budget

Stream temperature patterns are the result of the spatiotemporal variability in the energy available to the stream through advective and non-advective processes. The total energy available (Q_n) can be estimated as (Khamis *et al.*, 2015; Webb & Zhang, 1997):

$$Q_n = \Phi_{SW} + \Phi_{LW} + \Phi_{lat} + \Phi_{sens} + \Phi_{bed} + \Phi_{fric} + Q_{adv} \qquad (1)$$

where Φ_{SW}, the net shortwave radiation, is the energy coming directly from the sun (measured) reduced by reflection and shading (Glose, Lautz, Baker, 2017), and Φ_{LW} is the net longwave radiation, determined as the sum of two different contributions: longwave emitted from the atmosphere and landcover (Φ_{Lin}), and emitted longwave radiations at the stream surface (Φ_{emit}). These were either measured or estimated according to the commonly used Stefan-Boltzmann law (Glose *et al.*, 2017). The latent heat, Φ_{lat}, is the energy used for either condensation or evaporation estimated as a function of water temperature, wind speed and vapour pressure (Evans, McGregor & Petts, 1998; Webb and Zhang, 1997); Φ_{sens}, the sensible heat, is the transfer of energy between air and the stream surface and was calculated multiplying Φ_{lat} by the Bowen ratio (Garner, Malcolm, Sadler & Hannah, 2017); Φ_{bed}, the streambed conduction, is the heat transferred between the streambed and water calculated via Fourier's law using a shallow streambed temperature logger (Moore, Sutherland, Gomi & Dhakal, 2005b); Φ_{fric}, frictional heat, is the energy gained from the water as consequence of the friction against riverbed, estimated according to Theurer, Voos and Miller (1984). This approach provides a potential (maximum) value, as it considers that all of the potential elevation energy is converted into streambed friction. Finally, Q_{adv} represents the combined fluxes due to lateral or longitudinal advection. For our purposes, Q_{adv} is a lumped, unmeasured term whose cumulative effects on stream temperature are equivalent to the thermal influence of combined advective fluxes. We consider it

only qualitatively and relative to the non advective fluxes ($Q_{nadv} = Q_n - Q_{adv}$). For example, if stream temperature in a cross section is constant, Q_{nadv} must have the same magnitude of Q_{adv} but opposite sign. The supplementary material presents a more detailed description of the methods and equations utilized to estimate the heat fluxes. Hourly time series of non-advective fluxes were then derived and utilized to identify diel patterns.

3.2.3 Stream temperature patterns

Stream temperature time series were utilized to create stream temperature maps and boxplots, and to identify spatial and temporal patterns. Boxplots, providing a graphical representation of the main statistics of stream temperature distribution, allowed quantitative comparisons between different reaches and along the same reach. We considered sections 150 m apart and made a distinction between the warmer and cooler periods, early (June 22nd to August 26th) and late (August 27th to September 15th) summer, respectively.

3.2.4 Relationships between stream temperature and thermal processes

Linear regressions between mean daily air and water temperature were derived and used to investigate the relative impact of atmospheric thermal controls vs. other influences, and to quantify the 'stream thermal sensitivity from regression slopes and intercepts (Caissie, 2006; Kelleher *et al.*, 2012; Mayer, 2012). For example, streams dominated by advective (groundwater) fluxes are generally characterized by lower slopes and higher intercepts, while non-groundwater dominated catchments have higher slopes and intercepts closer to the origin (Caissie, 2006). Regression plots were derived both for entire reaches and for single sections.

Lag frequency distributions were utilized to qualitatively evaluate the importance of advective and non-advective fluxes on daily stream temperature maxima. We considered the difference between peaking time of air and stream temperatures (Lag T_{stream}- T_{air}), and non-advective fluxes

and stream temperature (Lag T_{stream}- Q_{nadv}) using hourly time steps. As peaking time depends also on upstream conditions, for each reach we considered stream temperature at the outlet as this represents an aggregate value for the entire reach. We hypothesize that temperature in streams more influenced by atmospheric conditions will closely follow air temperature and non advective fluxes and thereby exhibit a short lag, while those characterized by strong thermal influence from groundwater or other inflows will exhibit longer lags.

Finally, to complete our characterization of stream temperature dynamics in subarctic headwaters, we investigated the key drivers regulating overnight stream temperature, when shortwave radiation is not an important factor. In particular, we assessed the influence of air-water and bed-water interface processes, stream temperature, and discharge on nighttime stream temperature variations. For each component, coefficients of determination (R^2) were evaluated at each location. Air temperature was used as a proxy for air-water interface processes, while bed-water interface processes were derived as the sum of overnight conduction and friction fluxes. Nighttime was taken as 10 pm to 4 am when shortwave radiation input was near zero.

4 RESULTS

For the June – September 2016 study period, 187.4 mm of rain fell compared with a long-term average of 199 mm as measured at the BB weather station (Rasouli et al., 2019). The average temperature over this period was 9^{o}C, approximately 2.5^{o}C cooler than the long-term average, with all months having below normal values.

4.1 Stream hydrographs

The mean discharges for the study period (22 June – 15 September 2016) at the reach outlets of BB and GC were approximately equal at 1.6 and 1.9 mm d^{-1} (both ~0.13 m^3 s^{-1}), respectively.

Overall, the two headwaters exhibited similar temporal patterns in the hydrographs ($r = 0.74$), with a strong response to early season precipitation when the active layer was thin and frozen soils widespread, a moderate response in the middle of the summer when conditions were warmer and drier, and an increase in flow at the end of the season when precipitation increased and ET was low (Fig. 2a). Although flow patterns showed common features, significant differences were identified. GC exhibited a flashier response with greater variability (coefficient of variation, CV equal to 0.61 and 0.31 in GC and BB respectively) between wet and dry periods (peaks >6.0 mm d^{-1}, alternating to low baseflow <0.6 mm d^{-1}) than BB, which was characterized by a more persistent and stable flow (peaks <3.5 mm d^{-1}, baseflow >0.9 mm d^{-1}; Fig. 2a, 2c). The ratio between the specific discharges in GC and BB generally ranged between 0.5 and 1.5 (Fig. 2b). Early in the season, specific discharge in GC was much higher than in BB (ratio >1). In the middle of the summer (the driest period), around 12 July, this pattern switched, with BB having higher specific flow than GC (ratio <1). Towards the end of the season, the ratio progressively moved towards 1.

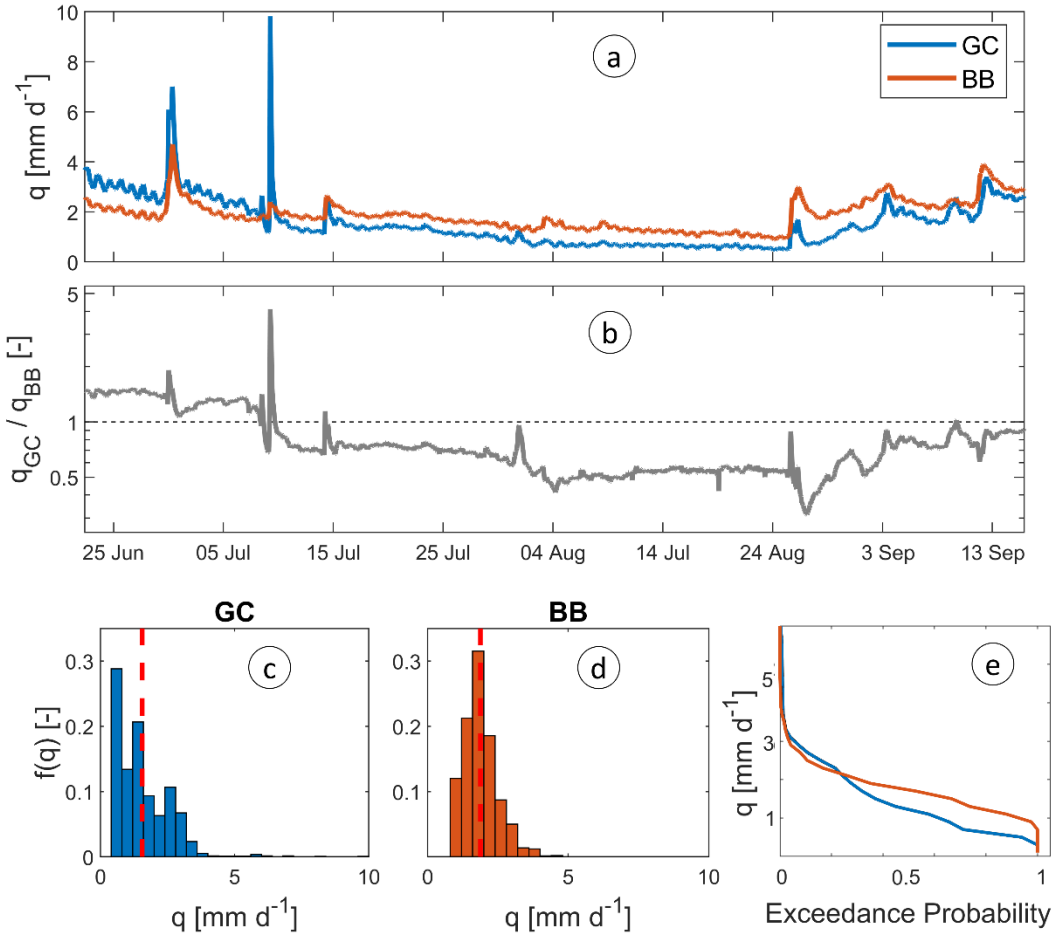
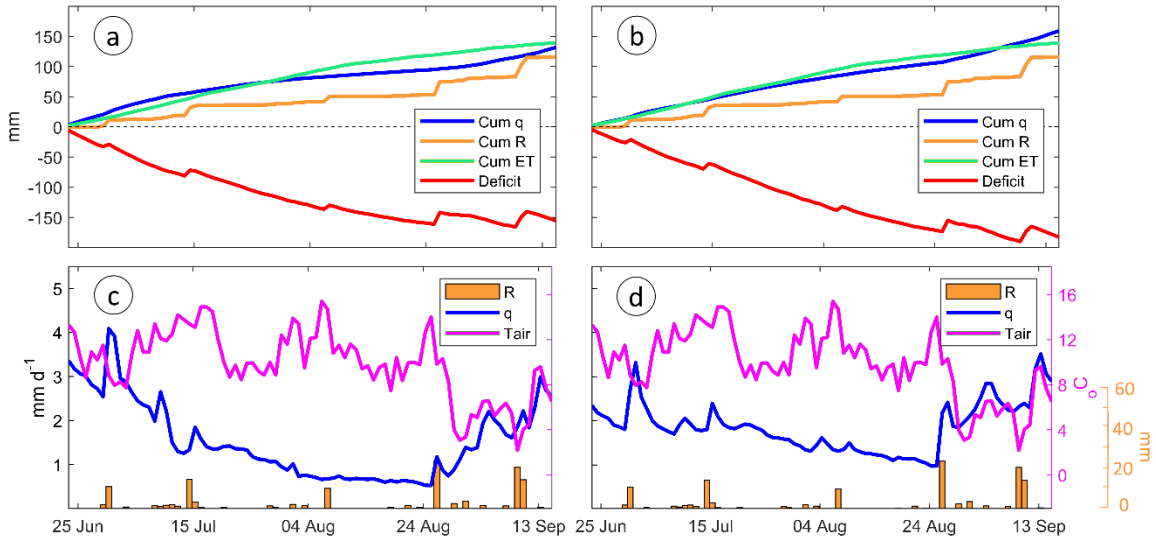


Fig. 2. GC and BB specific flow time series (a) and specific flow ratio (b, logarithmic scale). The highest divergence between the two sites occurred on 9 July when a major flood was recorded in GC but not in BB. Specific flow frequency distribution (interval = 0.4 mm d-1) in GC (c) and BB (d), and their exceedance probability curves (e). Red dashed line indicates the mean discharge.

The two sites also showed contrasting specific discharge frequency distributions. In BB, the discharge was more regular (CV = 0.31) and the distribution less skewed, with flows closer to the mean (Fig. 2d). Conversely, GC was characterized by a more erratic flow regime with enhanced intra-seasonal streamflow variability (CV = 0.50; Fig. 2c).

Fig. 3 a-b show the cumulative runoff (q), rain (R), evapotranspiration (ET), and deficit (P – (q + ET)) in GC and BB, respectively. Within the period considered, the loss of water is evenly divided between q and ET (average q/ET ratio ~1.04). In both sites, until late August (Fig. 3 c-d),

327 precipitation was not sufficient to compensate for water losses from q and ET. As a result, this
328 period was characterized by increasing water deficit and decreasing baseflow (Fig. 3 a-b). This
329 highlights the key role played by snowmelt and groundwater storage in sustaining summer
330 baseflow. In the latest part of the season, only when R increased and ET decreased (due to a
331 sudden drop in air temperature and declines in available energy, Fig. 3 c-d), the water deficit
332 stabilized and flow progressively increased.



333
334 Fig. 3. Cumulative specific flow (q), rainfall (R), evapotranspiration (ET), and deficit (P – (q +
335 ET)), for GC (a) and BB (b). Daily time series of flow, air temperature, and amount of
336 precipitation in GC (c) and BB (d).
337

338 ***4.2 Non-advective heat fluxes***

339 Figs. 4-6 show different aspects of non-advective fluxes in GC and BB. The contributions of
340 these fluxes were not constant and varied significantly both in time and space (between reaches
341 and sections within the same reach; Figs. 5-6). Overall, there is a clear diel pattern for all the
342 fluxes but the friction flux, whose variability in time is mainly discharge driven (Fig. 4a).

343
344 Net shortwave radiation (Φ_{SW}) was the primary mechanism of energy gain in both early and late
345 summer (Fig. 4 b-c, Fig. 6). Hourly averaged Φ_{SW} showed very high correlation (~0.97) with net

non-advective flux (Q_{nadv}; Fig. 4a). The magnitude and variability of Φ_{SW} significantly decreased in the last part of the season (Fig. 4 b-c). Medians decreased from ~104 to ~18 W m^{-2}, while the difference between the 3rd and the 1st quantiles was reduced by ~60 % (from ~309 to ~121 W m^{-2}). The contribution of Φ_{SW} ranged from 44 to 74 % of the total incoming flux (Fig. 5). Spatial variability was mainly driven by differences in shading (SF).

The second most important positive contribution was the (potential) bed friction flux (Φ_{fric}; Fig. 6) with a median of ~32 and ~47 W m^{-2} in early and late summer respectively (Fig. 4 b-c). The relative contribution of Φ_{fric} was higher in wetter periods (first and last part of the season; Fig. 6). Peaks of more than 40 % were reached at the end of the season when air temperature and shortwave radiation intensity were lower. Moving downstream, the relative contribution of Φ_{fric} tended to increase with discharge, as expected given the standard formulation (Theurer *et al*., 1984).

Overall, net longwave radiation (Φ_{LW}) provided a negligible positive contribution, as atmospheric and emitted longwave radiations had similar magnitude but opposite signs and tended to offset each other (Fig. 4a). More specifically, the contribution of Φ_{LW} was positive during the warmer part of the season (median ~7 W m^{-2}, Fig. 4b), and negative from late August onward (median ~ -7 W m^{-2}, Fig. 4c) when air temperature could account for up to 40 % of the total negative flux (Fig. 6).

Among non-advective fluxes, latent heat exchange (Φ_{lat}) was the dominant process by which heat was lost during the study period, with an average median of ~ -10 W m^{-2} (Fig. 4b-c). Its relative contribution to negative Q_{nadv} ranged between 33 and 100 % (Fig. 6).

Sensible heat flux (Φ_{sens}) was negative only for a few days, around 15 July in BB. Its contribution was significant in early summer (median ~27 W m^{-2}), and negligible in the late

season (median ~2 W m^{-2}; Fig. 4 b-c, Fig. 6). The relative contribution of Φ_{sens} did not exhibit relevant spatial fluctuations in GC or BB (Fig. 5)

Overall, bed conduction (Φ_{bed}) provided negative and positive contributions in early and late summer, respectively (median ~-4 and ~1 W m^{-2}; Fig. 4 b-c). Φ_{bed} was the second most important negative flux (Fig. 6) and the only flux that changed the sign of its contributions along the channel length, with the two streams showing specular patterns. In GC (BB) the flux was positive (negative) in the upper part and negative (positive) in the lower part of the reach (Fig. 5).

During daylight hours Q_{nadv} was always positive, but overnight its contribution could be either negative or positive (Fig. 4a). Average overnight Q_{nadv} was mainly controlled by air temperature, and thus by Φ_{LW}, Φ_{lat} and Φ_{lat} (R^2 respectively 0.37, 0.37, and 0.32; values averaged across reaches and sections).

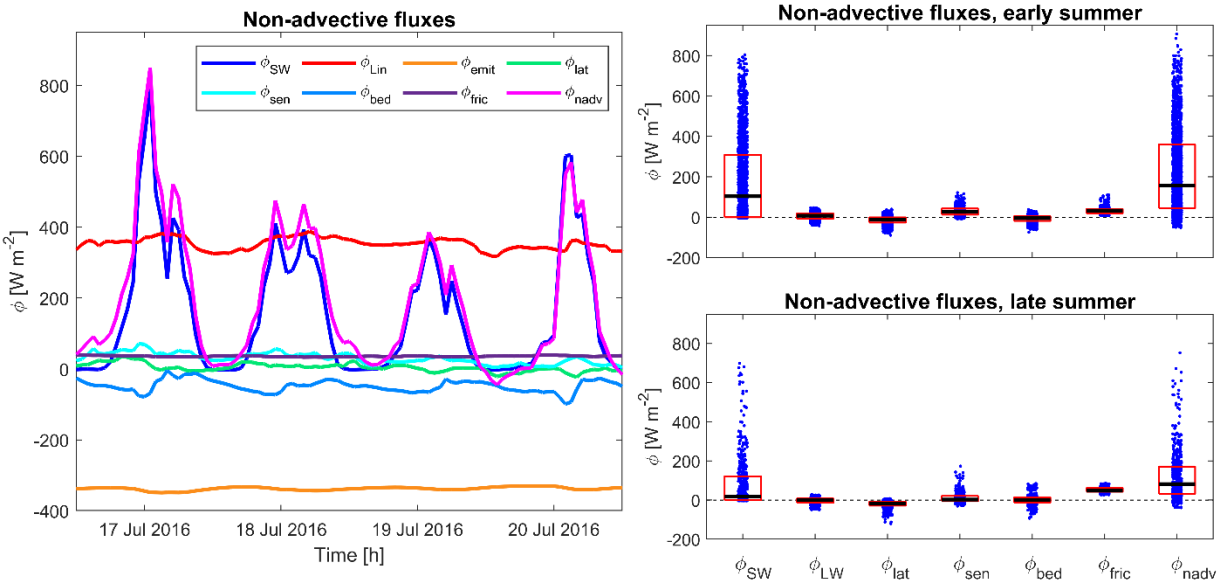


Fig. 4. Example of the multi-diel pattern of average (between GC and BB outlets) non-advective fluxes for a three-day period (a), and boxplots depicting the main statistic of the average (between sections and reaches) non-advective fluxes for early (22 June to 26 August, b) and late (27 August to 15 September, c) summer.

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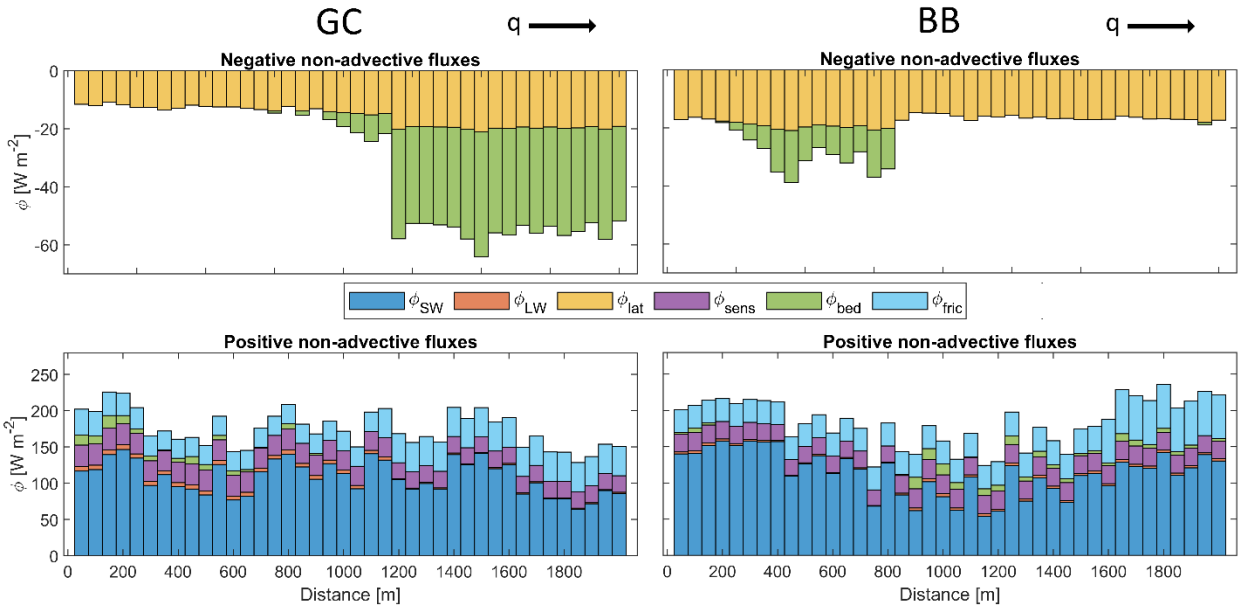
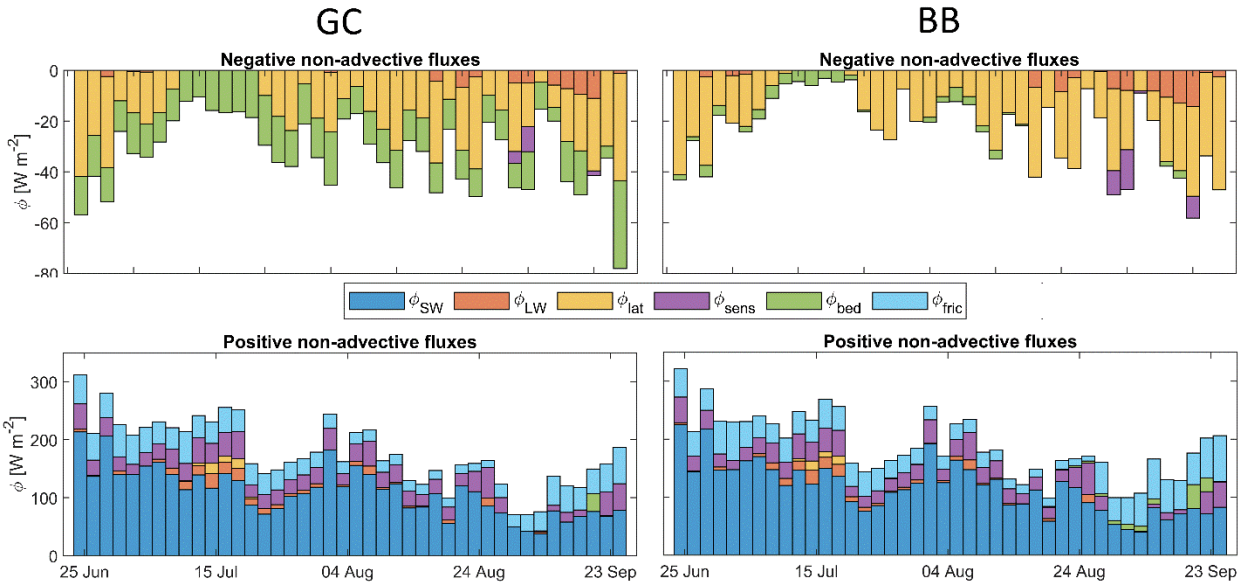


387
388 Fig. 5. For the two streams investigated, overall (averaged over the full study period) non-
389 advective negative fluxes and positive fluxes (below) vs. distance downstream from the upstream
390 station. Atmospheric observations did not vary spatially and for each stream an average
391 streambed temperature was considered. Each bar corresponds to a temperature monitoring
392 station.



393
394
395 Fig. 6. Average (over all stream locations) non-advective negative and positive fluxes (below)
396 vs. time for each of the two streams investigated. Fluxes were estimated at each stream
397 temperature monitoring location. Atmospheric observations did not vary spatially, and for each
398 stream an average streambed temperature was considered. Each bar corresponds to a 2-day
399 interval.

4.3 Thermal regimes

The observations captured both the warming and cooling periods over the summer season (Fig. 7). Stream temperature showed significant variability both in space and time at different scales. For the measurement period, average summer stream temperature in the two streams was ~3°C lower than average air temperature (9.0 °C). Stream temperature was slightly higher, but less variable in BB (mean 5.8 °C, average difference between 75^{th} - 25^{th} quantiles of 1.6 °C) than in GC (5.7 °C, 2 °C) (Fig. 8). For both sites, there was a decrease in stream temperature and variability between early and late summer, with an average median temperature decreasing 2.3 °C (from 6.0 to 3.7 °C) and 2.0 °C (from 6.2 to 4.2 °C) in GC (Fig. 8 a-c) and BB (Fig. 8 b-d), respectively. Generally, median stream temperature increased with distance downstream (Fig. 8); with the exception of BB in early summer, where medians were higher in the upper part of the reach (Fig. 8b). he upper tributary in GC (at 1150 m) was the only tributary significantly affecting stream temperature, with an average difference of ~0.6 °C between sections upstream and downstream of the confluence (Figs. 7-8). In early summer, it increased downstream temperatures, causing a reduction in variability (spread of blue dots in Fig. 8a), while in late summer, it decreased downstream temperatures without affecting variability (spread of blue dots in Fig. 8c).

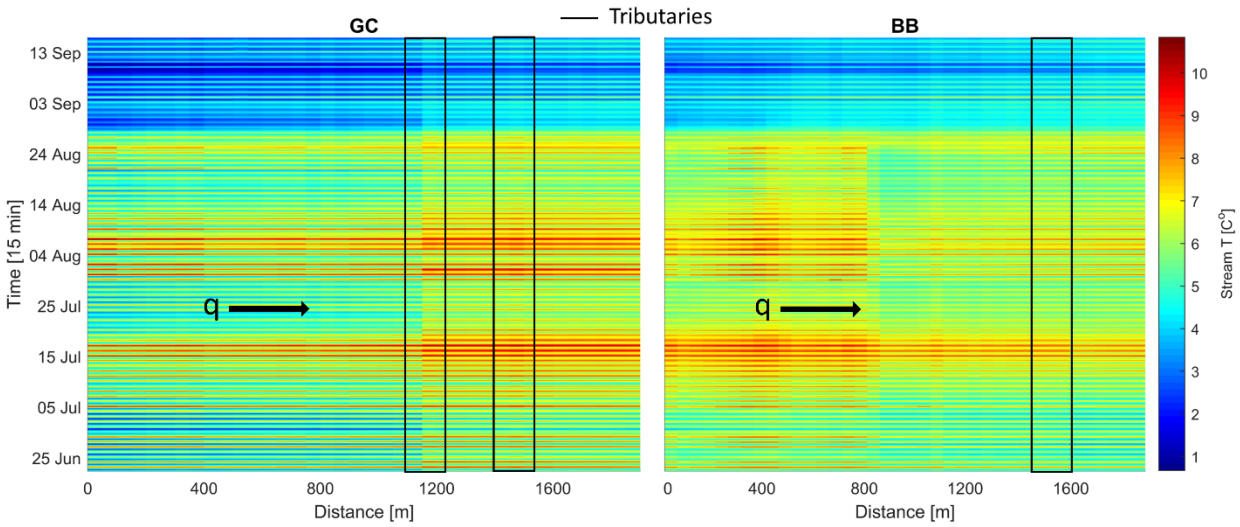


Fig. 7. Heat maps representing stream temperature observations in GC and BB reaches. Spatial and temporal resolutions are 50 m and 15 minutes. Black rectangles indicate the position of tributaries.

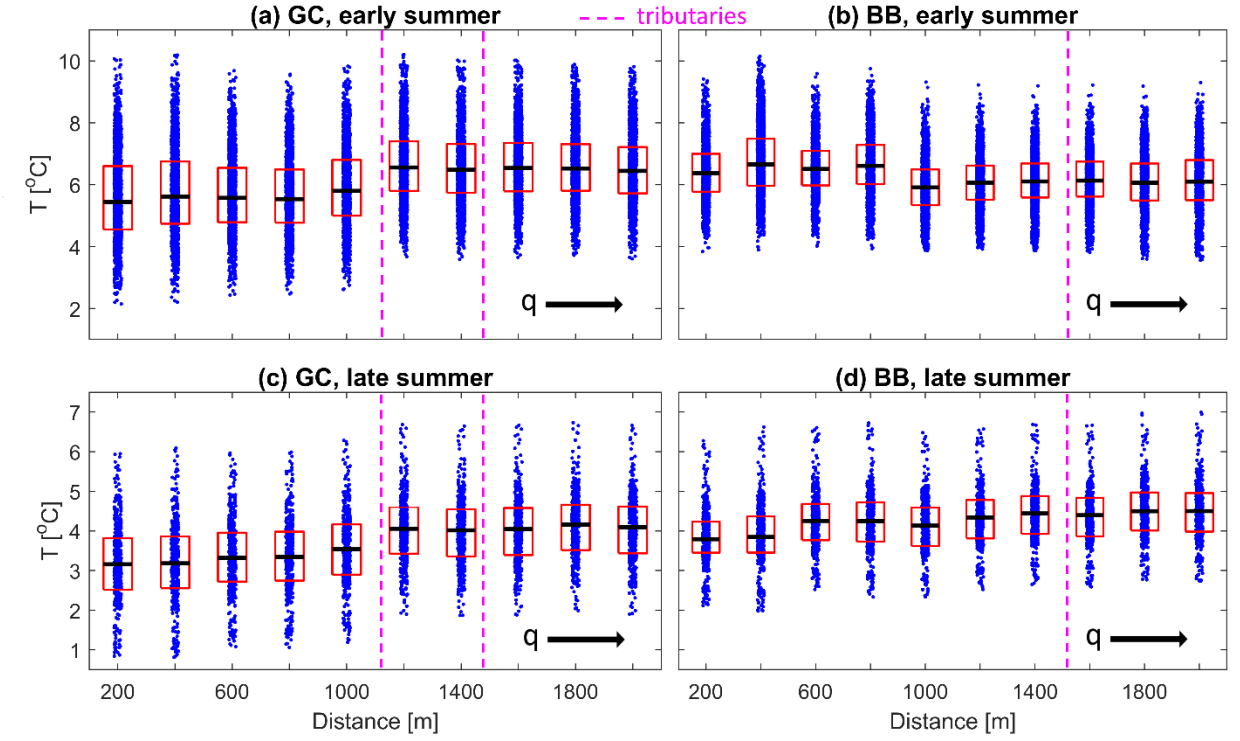


Fig. 8. Boxplots depicting some of the main statistics (medians and 25th and 75th quantiles) of temperature observations, early (22 June to 26 August) and late (27 August to 15 September) summer respectively, in GC (a, c) and BB (b, d). Blue dots represent single observations. Sections considered are 200 m apart.

4.4 Key controls on daily and sub-daily stream temperatures

Results from the regression analyses between mean daily air and stream temperature for the entire reaches yielded intercepts of 2.00 and 3.03 °C, and slopes (thermal sensitivities, Kelleher *et al.*, 2012) of 0.36 and 0.28 (both with a Newey-West Standard Error of ~0.022) in GC and BB, respectively (Fig. 9), suggesting a greater thermal influence of groundwater in BB than in GC. However, the slope and intercept were highly variable along the reaches of each stream (Fig. 10). In GC (BB) the slope ranged between 0.35 and 0.4 (0.22 and 0.39), with its standard error varying between 0.018 and 0.021 (0.017 and 0.025), while the intercept varied from 1.16 to 2.61 °C (2.13 to 3.56 °C). For both sites, the intercept followed an upward trend downstream. The slope in BB generally decreased moving downstream, while in GC, there was no clear spatial pattern.

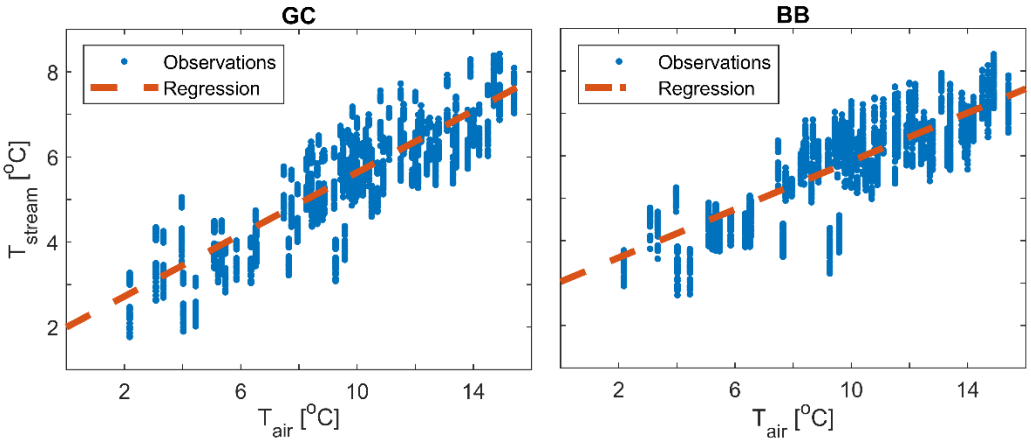


Fig. 9. GC (left) and BB (right), air vs. stream mean daily temperature and corresponding regression lines. All reach observations were included. The Newey-West Standard Error for the regression parameter are 0.0222 and 0.0224 for GC and BB.

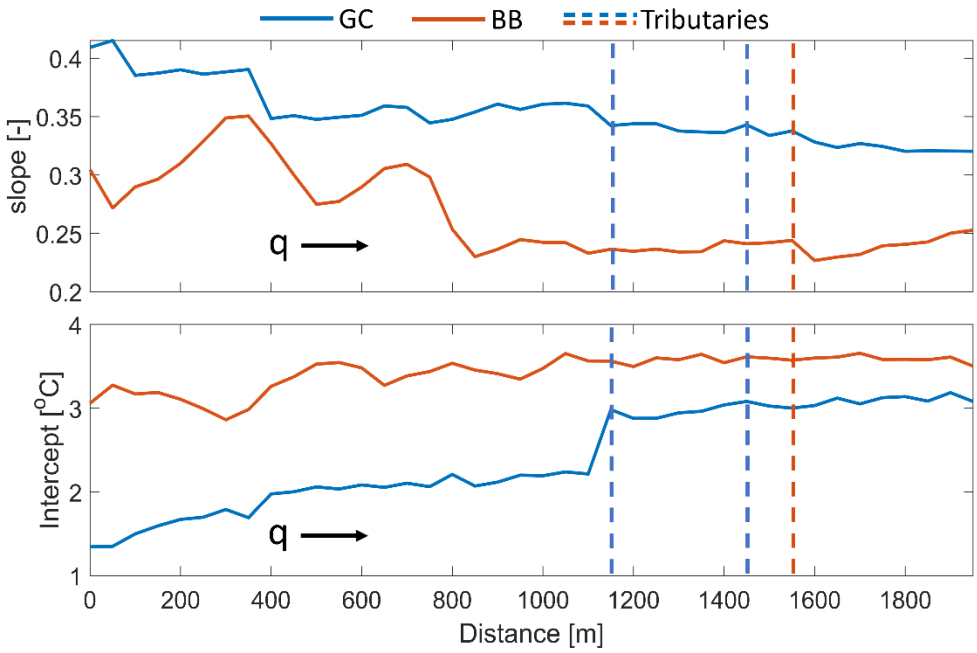


Fig. 10. Spatial variability of slope (upper) and intercept (lower) for linear regressions (mean daily air vs. stream temperatures) in GC and BB. Standard error of the regression parameter ranges between 0.018 - 0.021 in GC, and 0.017 - 0.025 in BB.

In both reaches, *Lag* Q_{nadv}-T_{stream} (average median equal to 1.5 h) was generally characterized by more positive values than *Lag* T_{air}-T_{stream} (average median equal to -0.5 h) (Fig. 11) as air temperature generally peaked 2 h after the non-advective flux (which as previously shown, was highly correlated to shortwave radiation). The 2 h delay was consistent in both sites.

Daily stream temperatures typically peaked later in GC than in BB. Negative values of the *Lag* Q_{nadv}-T_{stream} were unexpected for the summer season and, to our knowledge, are not reported elsewhere. The reasons for the slightly negative values (>-5 h) were: 1) the stream temperature, on a given day, did not exhibit a clear peak but a plateau where measurement errors and approximation might affect the timing of max Q_{nadv}; 2) the Q_{nadv} could be subject to strong, short time-scale fluctuations (e.g. due to variability in sky cover), and stream temperature peaks could occur at persistent high Q_{nadv} that did not coincide with or follow the air temperature daily peak. Alternatively, larger negative values <-7 h occurred when the overall energy balance was

negative throughout the day, as Q_{nadv} (in those days average peak ~165 W m^2) was not sufficient to compensate advective (negative, according to our definition) fluxes. Relatively low values of Q_{nadv} were caused either by very low solar radiation and/or decreases in daily air temperatures. Regarding the main drivers of nighttime stream temperature variations, our results showed that air-water interface heat fluxes were the most important explanatory variable with R^2 ranging between 0.32 and 0.45 and between 0.16 and 0.47 in GC and BB, respectively (Fig. 12a). R^2 varied spatially according to a clear spatial pattern where upward trends alternated to sudden but significant drops. Conversely, the other explanatory variables considered including discharge (Fig. 12b), average overnight stream temperature, and bed fluxes (conduction + friction) showed low correlation with overnight stream temperature variations (R^2 <0.16, <0.15 and <0.21 respectively).

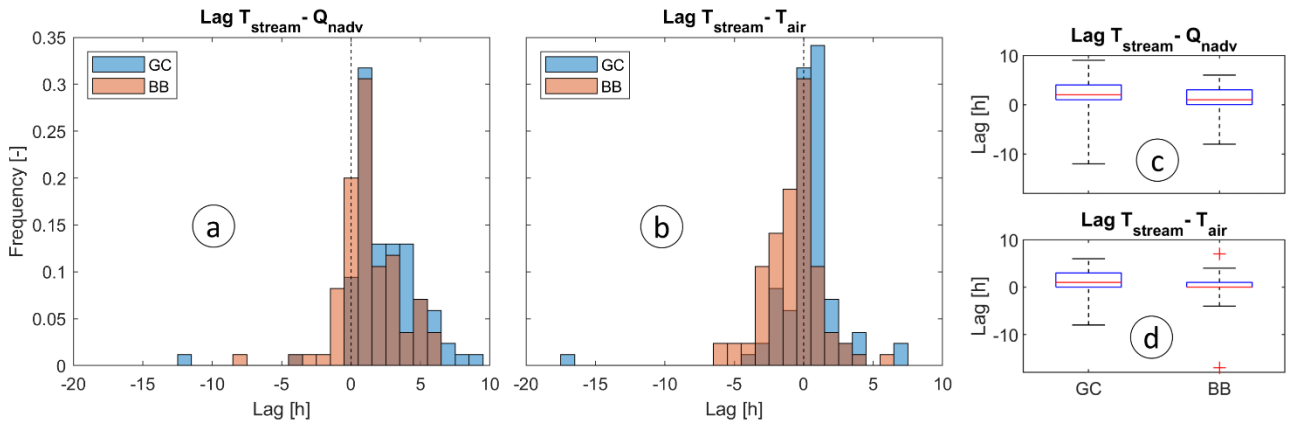


Fig. 11. Time lag [hours] distributions and boxplots between stream temperature and non-advective fluxes (a, c) and air temperature (b, d) in GC and BB.

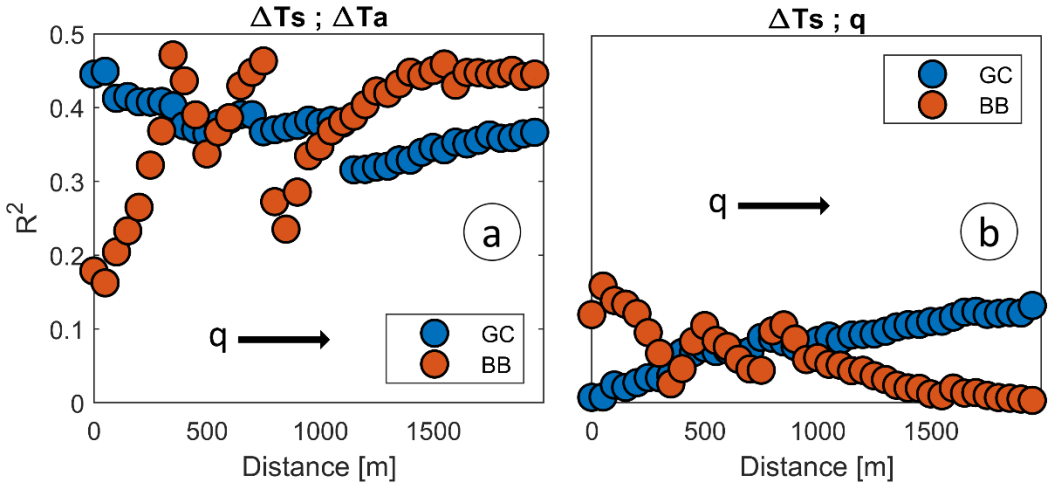


Fig. 12. Coefficient of determination (R2) between overnight stream (ΔTs) and air (ΔTa) temperature variations (a); ΔTs and specific discharge (q) (b) at the temperature monitoring sections in GC and BB. X axis indicates the distance from the upstream cross-section.

5 DISCUSSION

Stream temperature is the result of complex interactions between hydrological and thermal processes (Briggs *et al.*, 2018; Caissie, 2006; Vliet *et al.*, 2011). An empirical approach, based on high-resolution observations, was applied to provide insight on the hydrological and thermal regimes of two streams in adjacent permafrost influenced alpine headwater catchments (Fig. 13).

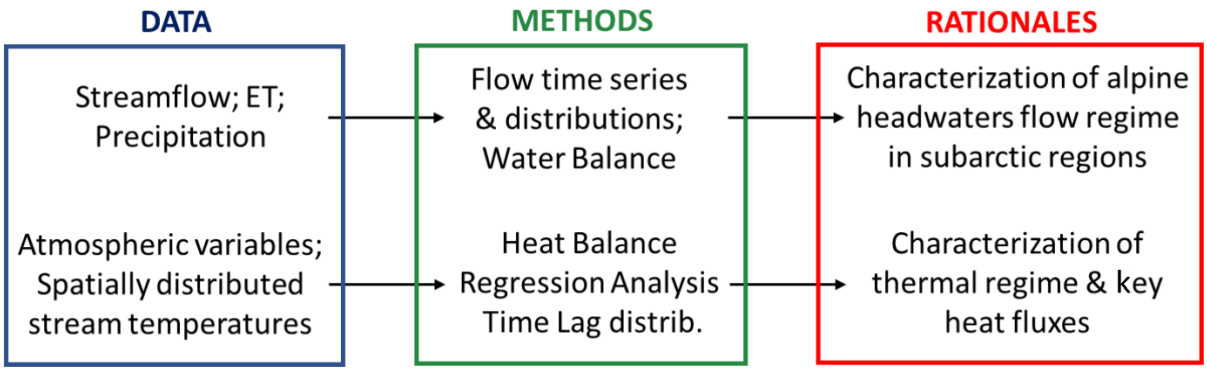


Fig 13. Flow chart reporting data, methods and rationales of the study.

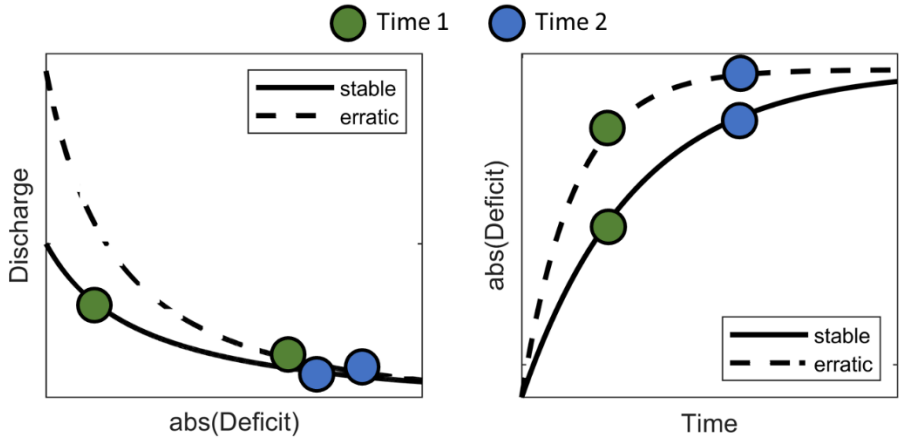
5.1 Hydrology of alpine headwater catchments

The two sites investigated had similar morphology, landscape, and exposure to climatic conditions. However, although they were characterized by similar mean specific discharge and correlated temporal flow patterns, the sites exhibited important differences. The flow regime in GC was more responsive to climatic variability and more exposed to extreme weather events (both peaks and troughs) than in BB. This could be due to a denser drainage network in GC, with several ephemeral tributaries activated during wet periods, that facilitates connectivity between hillslopes and the stream, reducing water residence time and limiting the buffering effects of catchment on snowmelt/precipitation signals (Hrachowitz, Soulsby, Imholt, Malcolm & Tetzlaff, 2010). Average summer specific discharge in GC and BB was more than three times higher than at the WCRB outlet (0.51 mm d^{-1}), as upland areas within WCRB are generally characterized by higher precipitation (approximately between 25 and 35 %) and less evapotranspiration than lower elevations (Pomeroy *et al*., 1998).

Both sites showed a daily and sub-daily flow response typical of headwater catchments, with short steep peaks alternating with more persistent low baseflow (Kosugi, Fujimoto, Katsura, Kato, Sando & Mizuyama, 2011; Penna, Meerveld, Zuecco, Fontana & Borga, 2016). Peaking times were generally synchronized, with random (evenly distributed between GC and BB) delays not greater than 1 h. Seasonal response was typical of alpine catchments dominated by permafrost at high latitudes, with the wettest period in late-spring/early-summer when most snowmelt occurs (Yang, Marsh & Ge, 2014; Ye, Yang, Zhang & Kane, 2009). Note that these observations occurred post-freshet when the streams were ice-free and full seasonal flows are reported in Shatilla and Carey (2019).

In northwest Canada, global warming has been identified as the main factor responsible for the positive annual and seasonal trends in air temperature between 1950 and 2012 (annual air temperature > 1.5 – 3 °C), trends that are likely to continue or accelerate in coming decades (Debeer *et al.*, 2016). According to predictions, an average increase in spring/summer air temperature between 2 to 3 °C is expected in the next 60 years (Debeer *et al*., 2016). In catchments where permafrost and frozen ground are important, such as GC and BB, baseflow in the dry summer period is largely driven by snowmelt processes (see high water deficit, Fig. 3) and is thus very sensitive to air temperature perturbations (Huntington & Niswonger, 2012). The increase in air temperature will hasten snowmelt, potentially leading to greater water deficits and lower flows later in the season. Counterintuitively, we believe that the more stable BB will be less hydrologically resilient to multi-decadal climate warming than GC, although the latter is more hydrologically responsive on diel and seasonal timescales. Fig. 14 shows a qualitative relationship between discharge and water deficit and how water deficit varies in time for a more hydrologically stable (e.g. BB) and a more responsive (e.g. GC) stream in a dry period with negligible precipitation. At first, under wet conditions, the erratic stream (dashed line) drains much faster than the stable stream (solid line). This will lead (in the erratic stream) to a more rapid increase in deficit and a substantial decrease in discharge. At a certain point the pattern would be inverted; discharge will become higher in the stable stream where the deficit will now start to increase more rapidly (green dots, current dry situation). At this stage, assuming an extended dry period, we expect a more rapid increase in water deficit and thus a greater reduction in flow in the stable stream (BB) rather than in the erratic one (GC) (blue dots, extended dry period due to early snowmelt). This is in agreement with what was found by Botter *et al.* (2013), who analytically demonstrated how streams characterized by unstable flow regimes

533 with enhanced intra-seasonal streamflow variability are likely to be more resilient to long-term
534 climatic variations. Changes in precipitation (both snow and rain) as a result of global warming
535 are also expected (Debeer *et al.*, 2016), and can either enhance or mitigate these effects.
536 However, predictions at a local scale are highly variable (both in space and time), and thus
537 difficult to quantify with confidence (Bates, Kundzewicz, Wu & Palutikof, 2008; Debeer *et al.*,
538 2016). Furthermore, in alpine permafrost catchments, the increase in precipitation can hardly
539 compensate the effect of an early snowmelt (Huntington & Niswonger, 2012), although in some
540 recent years the importance of rainfall events to sustain late summer flow both in GC and BB has
541 increased (Shatilla & Carey, 2019).



542
543 Fig. 14. Qualitative relation between discharge and deficit (left) and the variation of deficit in
544 time (right), considering no precipitation, for two streams with contrasting responses, stable
545 (solid line) and erratic (dashed line). Green and blue dots indicate values at an early and late
546 time, respectively.
547

548 ***5.2 Non-advective fluxes in alpine streams***

549 In agreement with previous studies located in headwater catchments, net radiation (shortwave +
550 longwave) was the flux providing most of the daily energy inputs (Moore, Spittlehouse & Story,
551 2005a; Moore *et al.*, 2005b; Webb & Zhang, 1997). The average proportion of approximately 60 %
552 net radiation was in agreement with that found in Hebert *et al.* (2011), but lower than the >70 %

found in other studies (e.g. Khamis *et al.*, 2015). A simple explanation is that most of those studies were set above the treeline (e.g. Khamis *et al.*, 2015) or subjected to bank clearcut (e.g. Moore *et al.*, 2005b), and thus had no significant shortwave reduction due to riparian vegetation shading (Garner *et al.*, 2017). Net radiation was mainly driven by shortwave radiation, while long-wave radiation was relatively negligible as incoming and emitted radiation tended to be of similar magnitude but opposite sign. This contrasts with what was found in the Kuparuk River (Alaska) where net long wave radiation was overall an important energy sink (King *et al.*, 2016).

Daily energy inputs with proportions of net radiation above 50 % in the early and middle part of the summer indicates that shortwave radiation remained the main positive contribution even on cloudy days (Khamis *et al*., 2015). In late summer, when daylight hours were reduced and discharge started to increase, the energy contributions of shortwave radiation and (potential) bed friction (~20 %) became comparable. As other studies in upland reaches suggested, the friction flux was a significant energy source both in GC and BB, with percentages between the 15.8 % (on average) estimated in 17 sites in the UK (Webb & Zhang, 1997; Webb et al., 2008) and the 30 % estimated in a headwater stream in Alaska (Chikita, Kaminaga, Kudo & Kim, 2010). In agreement with some previous studies, latent heat was the main cooling process in both stream reaches (Evans *et al.*, 1998; Hannah, Malcolm, Soulsby & Youngson, 2008; Hebert *et al.*, 2011), with its contribution being positive only for a few days during the warmest period (around 5 July). As reported by Khamis *et al.* (2015), sensible heat was overall an important energy source (~14 % on average), as air temperature was generally higher than stream temperature. In general, bed conduction had a buffering effect on diel and seasonal stream temperature peaks and troughs. As in Webb and Zhang (1997) and Evans *et al.* (1998), bed conduction was an important heat sink,

accounting for 30 % and 10 % of negative non-advective fluxes in GC and BB respectively, but a negligible source.

In alpine environments, change in vegetation type and distribution have been associated with recent global warming (Danby, Koh, Hik & Price, 2011; Grabherr, Gottfried & Pauli, 2010; Post *et al.*, 2009). In a study in the southwest Yukon Territory, Danby *et al.* (2011) concluded that in the last four decades, an increase in air temperature has led to an increase in shrub density. This suggests that in relatively narrow alpine headwaters such as GC and BB (width <3 m), with shortwave radiation dominating the non-advective fluxes, a change in vegetation (increase in density and height) could potentially buffer climate change effects on stream temperature (Moore *et al.*, 2005a, 2005b).

5.3 Thermal regimes in alpine headwater catchments

In both study reaches, our observations showed that stream temperature was very responsive to changes in diel climate conditions (see Fig. S1). Stream temperature showed high variability both in space and time, typical of low thermal capacity headwater streams during the summer season (Dick, Tetzlaff & Soulsby, 2015). Despite differences in magnitude, stream temperature variability between reaches and sections within the same reach was highly correlated ($r > 0.7$).

Stream temperature in BB was generally characterized by less variability (daily and sub-daily) than in GC. This is probably the result of a greater groundwater/discharge flow ratio (within the reach investigated), that could more effectively buffer non-advective flux troughs and peaks (Brown & Hannah, 2008; Dick *et al*., 2015; Johnson, Wilby & Toone, 2014; Snyder, Hitt & Young, 2015). Daily stream temperature typically peaked later in GC than in BB, suggesting that higher non-advective fluxes (as defined in section 3.2.2) are needed in BB to compensate for the negative contribution of advective fluxes and maintain an overall positive energy budget.

598 Regression analysis indicated that, in both sites (slope 0.36 and 0.28, and intercept 2.00 and
599 3.03 °C in GC and BB, respectively), summer stream temperature was less sensitive to
600 atmospheric conditions than in most of the 92 sites investigated in the western part of the US
601 (mean slope and intercept 0.45 and 0.74 °C) (Mayer, 2012). Nevertheless, caution is needed in
602 this type of comparisons as, at sub-annual time scales, the sensitivity of stream temperature to air
603 temperature may strongly be affected by the presence of clouds and the non-linear relationship
604 between air and stream temperatures at low (close to 0 °C) and high stream temperatures (Luce
605 et al., 2014). Notwithstanding, the use of weekly (instead of daily) data in Mayer (2012)
606 provided conservative results. In fact, as reported in Caissie (2006), different time scales yield
607 different air/stream temperature relationships: as the time scale increase (daily, weekly, monthly
608 and annually), slopes increase, and intercepts decrease.

609 Average stream temperature for the July – August period was about 4 °C lower than average
610 stream temperature measured at the WCRB outlet (9.8 °C) for the same period, and 10 °C lower
611 than the long-term (1975 - 2010) summer average measured near the Yukon River outlet (Yang
612 *et al.*, 2014). Although comparing annual and long-term averages might not be strictly
613 appropriate, the substantial increase in temperature between headwaters and large rivers
614 demonstrates the dominance of warming with respect to cooling processes during the summer
615 season. In agreement with other studies, this indicates the importance of small tributaries to
616 sustain biodiversification in large rivers by providing temperature discontinuities (Kiffney,
617 Greene, Hall & Davies, 2006; Brown & Hannah, 2008; Johnson *et al.*, 2014). It also affirms the
618 potential importance of alpine rivers for providing future cold-water climate refugia to sustain
619 cold-water species in the context of climate warming and future heat waves (Isaak *et al.*, 2015).

In GC, average overnight stream temperature tended to increase along the reach, ranging from 0.9 to 6.9 °C and 2.5 to 7.3 °C upstream and downstream, respectively. Conversely, in BB we saw a reduction of the range from 2.1 to 7.4 °C (upstream) and 2.9 to 6.7 °C (downstream) along the reach. The high correlation between overnight variations in air and water temperature suggests that each responded to the same temporal fluctuations in atmospheric inputs and thus that air-water interface processes are the dominant factors driving nighttime stream temperature (Johnson, 2004). The relationship was not constant in space but followed a cyclical pattern. This consisted of a sequence of upward trends, as the time of exposure of water to atmospheric factors increased along the stream, separated by sudden drops, likely due to localized surface (tributaries) and subsurface flows (preferential flow paths). Magnusson, Jonas and Kirchner (2012) investigated the sensitivity of overnight stream temperature variations to several explanatory variables in an alpine Swiss basin and found no significant relation between surface heat fluxes and overnight stream temperature dynamics. In their study, the head difference between channel and groundwater was identified as the most significant explanatory variable (not included in our study).

In the last two decades, the increase of stream temperature as a result of climate change, and its consequences on river ecosystems have been well documented (e.g. Isaak *et al.*, 2012). In addition, it has been shown that climate change effects on stream temperature in alpine headwaters where snowmelt is an important water source, are mainly driven by hydrological shifts in response to earlier onset of spring snowmelt (MacDonald *et al.*, 2014), a trend projected to continue in the future (Stewart, Morgenstern & McDonnell, 2010). Our results suggest that, due to its more stable and resilient thermal regime, BB will be less affected by direct climate change effects. The relatively high contribution of groundwater, generally with opposite sign

with respect to net non-advective fluxes, can effectively buffer extreme climatic conditions. This occurs as groundwater input tends to moderate both temperature peaks and troughs, but its ability to attenuate future low frequency warming may be impacted by future aquifer warming (Kurylyk *et al.*, 2014). However, in permafrost sites, discharging groundwater may stay cool in a warmer climate as it continues to thermally interact with ground ice. Nonetheless, BB seems to be more exposed to indirect effects, as in BB a greater reduction in discharge and thus in thermal capacity is expected (section 5.1). In conclusion, we believe long-term studies, meaning long-term hydrologic and atmospheric data, are necessary in order to make reliable long-term predictions accounting for global warming, given the complexity of the thermal and hydrological processes involved.

5.4 Study limitations

The uncertainty associated with measuring instruments was deemed negligible compared to the magnitude of the variations in space and time considered and did not significantly impact our results. In some specific occasions, small-scale spatial variability in atmospheric conditions and localized rainfall events may not have been well captured by a single weather station. A critical analysis of the data helped to identify such events. For example, on 9 July a major flood was recorded in GC (confirmed by flow measurements in upstream gauges), while precipitation and flow records in BB showed only a mild event. Nevertheless, we believe that the overall consistency of our results demonstrated that these errors did not impact our analysis and findings significantly.

A further source of uncertainty is associated with errors in flow estimations, the sources of which are well documented in the literature. Despite the lack of a control structure, repeated measures at GC over 20 years using both dilution and velocity-area methods across a range of flow

indicates an error < 10%, which is similar to that reported by Hood, Roy and Hayashi (2006) using the velocity-area method for several alpine streams. While we do not have long-term flows from BB, similar methodology and a robust rating curve in 2016 suggests that the error in flow estimates is significantly lower that the differences in specific flows for the two streams.

As a single weather station was utilized, it was not possible to account for small-scale variability in some of the atmospheric variables. It is important to note, however, that many stream temperature studies have relied on meteorological stations several kilometers away, especially for streams that are logistically difficult to reach (Hebert et al., 2011). The most affected atmospheric variables include wind speed and humidity, for which we could not account for riparian vegetation effects. This might result in estimating potential (maximum in absolute terms) Φ_{lat} and Φ_{sens} rather than actual values.

Another important limitation that future studies should try to overcome is related to the timespan considered. The study provided useful insights on the thermal response of streams to hydrologic and atmospheric processes; however a short-term study such as ours cannot be confidently employed to identify any long-term trend or assess climate change impacts (Bolduc & Lamoureux, 2018), revealing the importance of long-term hydrologic and thermal monitoring (e.g. Rasouli *et al.*, 2019).

Furthermore, our study is limited to the summer season as it was logistically difficult to collect observations during the colder period of the year. This study period is common among stream temperature studies set in remote regions with challenging climatic conditions such as ours, often preventing a full characterization of the hydrological and thermal cycles across seasons. Thus, to fill this gap, future studies should attempt to include winter stream thermal regimes, as these have been shown to also be important for aquatic habitat (e.g., Cunjak & Power, 1986).

5.5 Conclusions and future directions

The present study provided insights on the processes governing the hydrologic and thermal regimes in two adjacent subarctic alpine headwaters. This is important as alpine headwaters in permafrost regions have been shown to be among the most sensitive environments to climate change. Our results highlighted the major role played by groundwater on the hydrologic and thermal response of sub-alpine headwater catchments. A high groundwater/discharge ratio could effectively moderate both flow and stream temperature extremes. Overall, in both streams, solar radiation and latent heat were respectively the most important heat source and sink.

The last decade has seen a reduction of experimental long-term data collection (Tetzlaff, Carey, McNamara, Laudon & Soulsby, 2017). In this regard, we want to highlight the importance that ongoing collection of field data has on supporting both empirical and model-based studies. In addition, in headwater catchments at high latitudes or elevations, long-term and spatially distributed studies are extremely rare (e.g. Bolduc and Lamoureux, 2018). Therefore, we believe that in future studies, long-term data sets (in alpine headwater catchments) in combination with models should be used to: (1) identify significant long-term trends; (2) quantify the potential effects of changing climate and land use (e.g. change in vegetation related to global warming) on flow and thermal regimes; and (3) provide specific tools to assist catchment managers to evaluate and apply suitable management strategies.

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