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Quantifying the contribution of glacier runoff to streamflow in the upper Columbia River basin, Canada

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Abstract

Glacier melt provides important contributions to streamflow in many mountainous regions. Hydrologic model calibration in glacier-fed catchments is difficult because errors in modelling snow accumulation can be offset by compensating errors in glacier melt. This problem is particularly severe in catchments with modest glacier cover, where goodness-of-fit statistics such as the Nash-Sutcliffe model efficiency may not be highly sensitive to the streamflow variance associated with glacier melt. While glacier mass balance measurements can be used to aid model calibration, they are absent for most catchments. We introduce the use of glacier volume change determined from repeated glacier mapping in a guided GLUE (generalized likelihood uncertainty estimation) procedure to calibrate a hydrologic model. We also explicitly account for changes in glacier area through the calibration and test periods. The approach is applied to the Mica basin in the Canadian portion of the Columbia River basin using the HBV-EC hydrologic model. Use of glacier volume change in the calibration procedure effectively reduced parameter uncertainty and helped to ensure that the model was accurately predicting glacier mass balance as well as streamflow. The seasonal and interannual variations in glacier melt contributions were assessed by running the calibrated model with historic glacier cover and also after converting all glacierized areas to alpine land cover in the model setup. Although glaciers in the Mica basin only cover 5 % of the watershed, glacier ice melt contributes up to 25 % and 35 % of streamflow in August and September, respectively, and is particularly important during periods of warm, dry weather following winters with low accumulation and early snowpack depletion. The approach introduced in this study provides an effective and widely applicable approach for calibrating hydrologic models in glacier fed catchments, as well as for quantifying the magnitude and timing of glacier melt contributions to streamflow.

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1 Introduction

In many mountainous regions, glacier melt makes significant contributions to streamflow, particularly in late summer during periods of warm, dry weather (Stahl and Moore, 2006; Kaser et al., 2010). Understanding the quantity and timing of these contributions is important for a range of purposes, including short-term forecasting of reservoir inflows and long-term projections of the potential hydrologic effects of climate change. This knowledge is particularly critical given that these contributions are likely to decrease as glaciers retreat (Stahl et al., 2008; Huss et al., 2008; Rees and Collins, 2006; Marshall et al., 2011), with implications for both water resources management and aquatic ecology (Milner et al., 2009; Moore et al., 2009).

In catchments where glacier mass balance and snowline observations exist, a water balance approach can be used to estimate glacier contributions to streamflow (Young, 1982). Alternatively, empirical analysis of the contrasting responses of glacier-fed and unglacierized catchments can provide insight (Collins, 1987; Stahl and Moore, 2006). Deterministic hydrologic models can be used to quantify glacier melt contributions to streamflow (e.g., Comeau et al., 2009; Schaefli and Huss, 2011). However, the presence of glaciers can exacerbate problems with equifinality in streamflow modelling, as an incorrect simulation of snow accumulation can be offset by compensating errors in the simulation of glacier melt. Problems with equifinality can be reduced by constraining a model to reproduce glacier mass balance or equilibrium line altitudes in addition to streamflow (Braun and Aellen, 1990; Moore, 1993; Schaefli et al., 2005; Stahl et al., 2008; Konz and Seibert, 2010; Schaefli and Huss, 2011). Unfortunately, glacier mass balance observations are not available for most catchments. In cases where mass balance data exist, they may be of limited value in large catchments with substantial climatic gradients, due to difficulties in extrapolating from a single glacier or even a small number of monitored glaciers.

Most modelling studies have focused on catchments with substantial glacier cover, in excess of 10 % of the catchment area. However, Stahl and Moore (2006) found that the effects of glacier cover on late-summer streamflow can be detected in catchments with as little as 2 to 5 % glacier coverage. Equifinality may be especially problematic in large catchments with modest glacier cover (less than 10 %) given the relatively small variance in streamflow associated with glacier melt contributions.

Another challenge in modelling glacier melt contributions to streamflow is that glacier area and volume can vary significantly over periods of a decade or more. Most hydrologic models, however, do not account for changes in area or volume through time. Treatment of glaciers as static during a calibration period could result in distortion of the parameters that control snow accumulation and glacier melt, which would then result in biased future projections. While some studies have accounted for transient changes in glacier cover in future projections (Rees and Collins, 2006; Stahl et al., 2008; Huss et al., 2008), we are unaware of any studies that include changes in glacier cover in the model calibration and testing process.

The objective of this study was to develop an approach for calibrating hydrologic models in large catchments with modest glacier cover and no mass balance observations, and to use the model to characterise the magnitude and timing of glacier melt contributions to streamflow, along with an assessment of uncertainty. This study used glacier volume and area changes throughout the basin to assist in calibration, which were computed by analysing digital elevation models (DEM) and maps of glacier cover (Schiefer et al., 2007).

2 Methods

2.1 Study area

The study focused on the Mica basin, a major tributary to the Canadian portion of the Columbia River basin in British Columbia (Fig. 1) with a drainage area of 20 742 km^2. Elevation in the basin ranges from 579 m above sea level (a.s.l.) at Mica dam (MCA in Fig. 1) to 3685 m a.s.l. In 1985, glaciers covered 1268 km^2 in the Mica basin, representing 6.1 % of the total basin area. Between 1985 and 2000, the glacier area decreased by 101 km^2, and an additional 80 km^2 of glacier area was lost between 2000 and 2005, thus reducing glacier cover to 5.2 % of the basin area. About 50 % of the basin consists of open land cover types (i.e., alpine areas, range lands, agricultural lands, recently logged areas), and about 45 % of the area is forested.

Mica dam is one of the largest earth-filled dams in the world and impounds Kinbasket Reservoir. The Mica project ranks third for generating capacity in BC Hydro's energy portfolio. For BC Hydro, the crown corporation that operates the Mica project and other major hydroelectric facilities in the Columbia River basin, quantifying glacier influences on reservoir inflow is an integral part of its water resource management approach, which aims to maximize triple bottom line outcomes. This study is particularly motivated by the upcoming review of the Columbia River Treaty in 2014 and the associated need to understand how reservoir inflows may change in response to climate change.

2.2 Data

Data from five climate stations within or just outside Mica basin were available for modelling (Fig. 1). Mica dam climate station (MCA) has the longest climate record, dating back to 1965. Rogers Pass climate station (RGR) data start in 1967, Radium climate station (RAD) in 1969, Molson Creek climate station (MOL) in 1986, and Floe Lake climate station in 1993. Backfilled climate data were needed to calculate historic changes in streamflow. To extend the records for all climate stations back to 1965, we computed factors for each three-month quarter to rescale precipitation data from MCA. Air temperature was estimated based on linear regressions for each quarter of the year. Only measured climate data were used for model calibration and testing, except for eight years of backfilled data from FLK (1985–1993).

Streamflow data are inflows to Kinbasket Reservoir computed from a water balance based on the rates of release through the dam and changes in water level. The data were provided by BC Hydro at a daily resolution. Although evaporation from the

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reservoir is not included in the computed inflows, estimates based on reservoir area and potential evaporation indicate it should not exceed about 1 % of inflow.

Snow water equivalent (SWE) data for three snow pillows located at the FLK, MOL, and RGR climate stations (Fig. 1) were available from 1995 onwards. Glacier coverages were derived from Landsat Thematic Mapper scenes for 2005 and 2000 and from high altitude, aerial photography for 1985 (Bolch et al., 2010). Glacier volume loss was calculated from digital elevation models (DEM) derived from the 1999 Shuttle Radar Topography Mission (SRTM) and from aerial photographs taken between 1982 and 1988, which have a median weighted date for Mica basin of 1985 (Schiefer et al., 2007). The estimated ice volume loss from 1985–1999 was 7.75 km^3 with a specific thinning rate of 0.43 $m\,yr^{-1}$. Taking mapping uncertainty into account, ice volume loss from 1985–1999 lies between 6 and 9 km^3.

2.3 The HBV-EC hydrologic model

The HBV-EC model is a Canadian variant of the HBV-96 model (Lindstrom et al., 1997). It has been incorporated into the EnSim Hydrologic modelling environment (now known as Green Kenue) (Canadian Hydraulics Centre, 2006). The ability of HBV-EC to provide accurate predictions of streamflow in British Columbia's mountain catchments was demonstrated in an intercomparison study of watershed models for operational river forecasting (Cunderlik et al., 2010; Fleming et al., 2010). The model algorithms have been described in detail by Hamilton et al. (2000), Canadian Hydraulics Centre (2006) and Stahl et al. (2008). Key features are presented below.

To minimize computational effort, HBV-EC is based on the concept of grouped response units (GRUs), which contain grid cells having similar elevation, aspect, slope, and land cover. HBV-EC can model four land-cover types: open, forest, glacier and water. To represent lateral climate gradients, HBV-EC allows for subdividing a basin into different climate zones, each of which is associated with a single climate station and a unique parameter set. Water draining from non-glacier GRUs is routed through two lumped reservoirs representing "fast" and "slow" responses. To predict the discharge

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for a given time step, HBV-EC sums output from the two non-glacier reservoirs and the reservoirs associated with glacier GRUs (see below).

The temperature-index-based snow melt algorithm from HBV-96 was adapted by Hamilton et al. (2000) to account for the effects of slope, s, aspect, a, and forest cover. In HBV-EC, daily snowmelt (M) (mm day^{-1}) is calculated from daily mean air temperature (T_{air}) (°C) as follows:

$$M(t) = C_0(t) \times \mathrm{MRF} \times (1 - AM \times \sin(s) \times \cos(a)) \times T_{\mathrm{air}} \quad (1)$$

where C_0 is a base melt factor (mm day^{-1} °C^{-1}) that varies sinusoidally between a minimum value (C_{min}) at the winter solstice to a maximum value at the summer solstice ($C_{\mathrm{min}} + DC$) to account for seasonal variations in solar radiation, and DC is the increase in melt factor between winter and summer solstices. The melt ratio for forests, MRF, ranges between 0 and 1 and reduces melt rates under forests compared to melt at open sites. The coefficient AM controls the sensitivity of melt rates to slope and aspect. For glacier GRUs, melt is computed as for an open site (MRF = 1) until the previous winter's snow accumulation has ablated. At that point, glacier melt is computed by multiplying open site melt by the coefficient MRG, which typically ranges between 1 and 2, to reflect the reduction in surface albedo.

Storage and drainage of meltwater and rain for each glacier GRU are modelled using linear reservoirs. The outflow coefficient for each GRU depends on snow depth, ranging from a low value (KG_{min}) when the GRU has deep snow cover to a maximum value ($KG_{\mathrm{min}} + dKG$) when the GRU is snow-free (Stahl et al., 2008). This representation accounts for seasonal changes in the efficiency of the glacier drainage system. Glacier mass balance can be computed by post-processing HBV-EC model output for the glacier GRUs. Glaciers in HBV-EC cannot vary in area or volume during a model run without stopping and restarting the simulation.

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2.4 Calibration and testing

The model was calibrated for the period 1985–1999, the same period for which the glacier volume loss was calculated. Calibration runs were split into two time periods, each with a five-year spin-up period to ensure that storages in the model, in particular the slow reservoir storage, equilibrated with the forcing data. Simulations for the first period, 1985–1992, used the 1985 glacier coverage, while the second period, 1992–1999, was based on glacier coverage from 2000. This approach explicitly incorporates the hydrologic effects of glacier retreat during the calibration period, and helps to avoid parameter bias that could arise from using only one (static) glacier coverage. Glacier net mass balance (b_n) for the entire basin was derived from net mass balances for each GRU and compared to geodetically calculated glacier volume loss.

The period 2000–2007, with glacier cover based on data from 2005, was used as an independent test period. Model predictions were compared to observed streamflow data and SWE data from the three snow pillow sites (FLK, MOL, RGR, in Fig. 1). Since HBV-EC does not predict state variables for a specific location but only for each GRU, observed SWE data were compared with simulated SWE from the GRUs in which the snow pillows are located. The SWE data were not used in model calibration, and thus represent an independent test of the model.

2.5 A "guided" GLUE approach to address parameter uncertainty

A common approach to address uncertainty in model predictions is to generate random samples from the usually high-dimensional parameter space and subsequently to pick the best performing parameter sets according to one or multiple criteria (e.g. Konz and Seibert, 2010; Stahl et al., 2008 for glacier related applications). However, in a high-dimensional parameter space, random sampling with even thousands of model runs does not guarantee that the "best" parameter combinations are found. Without prior knowledge of how well the "best" possible solution performs, the modeller will usually relax criteria in order to obtain enough acceptable parameter sets, with the possible

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result that criteria for acceptable parameter sets are more relaxed than necessary. To ensure that the final ensemble parameter set contains solutions close to the "best" possible solution(s) within a parameter space, we modified the Generalized Likelihood Uncertainty Estimation (GLUE) approach outlined in Beven and Freer (2001) and Freer et al. (1996) to an approach that can best be described as a "guided" GLUE approach (Fig. 2).

The calibration procedure in this approach starts with finding a benchmark parameter set by maximizing the Nash-Sutcliffe efficiency (E) or, in terms of GLUE, the generalized likelihood measure. This was done with *genoud* (Mebane and Sekhon, 2009), an optimization algorithm in R (R Development Core Team 2009) that combines evolutionary algorithm methods with a steepest gradient descent algorithm. A large negative number was returned for parameter sets that did not fulfill the multiple criteria listed in Fig. 2 to ensure that the optimization algorithms not only maximized E but also searched for solutions that met the additional criteria. In a second step, a Latin Hypercube Search (LHS) with 10 000 model runs was performed. Latin hypercube designs are most often used in high-dimensional problems, where it is important to sample efficiently from distributions of input parameters. Parameter sets from the 10 000 model runs were constrained by criteria given in Fig. 2. If no parameter sets with Nash-Sutcliffe efficiencies greater than the benchmark efficiencies minus a threshold were found, all parameter sets were rejected. There are two ways to proceed when no parameter sets are found by the LHS: either increase the sample size or adjust the prior parameter distributions (decrease the range). Increasing the sample size is the favored solution because it should lead to a more diverse set of parameters. However, the number of model runs is limited by computational power (even with multiple CPUs it would take months for the Mica basin). Given time constraints, we chose to adjust the prior parameter distributions. With adjusted (narrowed) parameter ranges, the LHS was repeated until enough parameter sets (~20–30) were found that fulfilled all criteria (i.e., "behavioral" parameter sets in GLUE terminology). The parameter ranges for model calibration and uncertainty analysis were based on default values provided in the HBV-EC manual (Canadian Hydraulics Centre, 2006), values reported in previous studies (Stahl et al., 2008; Hamilton et al., 2000), the authors' experience with applying HBV-EC on other catchments, and by visually testing the influence of parameters on the simulated hydrograph. With the modest glacier cover in Mica, a visual inspection of simulated hydrographs provides more information on the sensitivity of modelled streamflow to the various glacier parameters than a single goodness of fit measure such as E. Prior parameter distributions for LHS were assumed uniform at all stages.

2.6 Modelling the contributions of glacier ice melt to streamflow

An estimate of the contribution of glacier ice melt to discharge and the associated uncertainty was calculated as the difference between streamflow simulations with and without glaciers for each ensemble member. In the no-glacier runs, all glacier cover was converted to open land cover to account for the fact that snowmelt and rainfall runoff from the areas currently covered by glaciers would occur even if the glaciers completely disappeared.

To accommodate changes in the glacier extents and elevations through time, HBV-EC was run using scripts that would update the glacier GRUs used in the simulations based on the observed glacier extents in 1985, 2000, and 2005. The updating involved stopping the simulation, reading in the new glacier extents, updating the definitions of Grouped Response Units and state variables, and then continuing the simulation, including a five year spin-up period. Transient runs from 1972 to 2007 were obtained by running HBV-EC from 1972 to 1992 with the observed 1985 glacier cover, from 1993 to 2000 with the glacier cover from 2000, and from 2001 to 2007 with the observed 2005 glacier cover. The assumption that glacier areas did not change appreciably from 1972 to 1985 is supported by physically based distributed modelling of glacier dynamics with the UBC Regional Glaciation Model (Garry Clarke, University of British Columbia, unpublished results).

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3 Results

3.1 Model calibration and uncertainty analysis

The benchmark parameter set obtained by the combined evolutionary-steepest gradient optimization matched observed the streamflow data with E of 0.93 for the calibration period (1985–1999) and 0.95 for the test period (2000–2007). A first 10 000 run LHS within the initial parameter ranges (parameter range step 1 in Fig. 2) found no acceptable parameter sets that met all criteria. Although 28 parameter sets had $E > 0.91$, all of these parameter sets were rejected because none fulfilled all of the additional criteria. In the absence of prior knowledge of the benchmark E, the common procedure would now have been to find acceptable solutions with relaxed criteria. However, the benchmark parameter set indicates that there are better performing solutions within the initial parameter space; 10 000 runs are too few to sample the parameter space for acceptable solutions. A second LHS with adjusted parameter ranges found 17 acceptable parameter sets, but histograms indicated that two parameters in the acceptable parameter sets were predominantly sampled near a range boundary and therefore a third LHS with slightly refined parameter ranges was performed. From the 10 000 model runs in the third LHS, 705 parameter sets met the final criterion of $E > 0.92$, but only 23 of these also met the additional criteria (Fig. 2).

The calibrated parameters with the highest correlations to glacier net mass balance (b_n) and E are temperature lapse rate (T_{lapse}), melt factor at winter solstice (C_min), and precipitation lapse rate (P_lapse) (Table 1). Other important parameters are the ratio between melt rates for glacier ice and seasonal snow (MRG) and the increase of melt factor between winter and summer solstice (DC). MRG and DC are both correlated with b_n at all steps during the uncertainty analysis, but are correlated with E only in the first LHS with wide parameter ranges. The routing parameters with the highest correlation with E are the fast reservoir release coefficient (K_F) and the fraction of runoff directed to the slow reservoir (FRAC). The exponent to adjust the linearity of the release rate of the fast reservoir, α, has little influence on E. Glacier reservoir coefficients and the

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melt reduction factor under forest (MRF) show weak correlation to both glacier volume change and E.

A wide range of modelled glacier volume changes can lead to values of E close to the benchmark (Fig. 3). Results from the first LHS suggest that equifinal parameter solutions are possible with glacier volume losses ranging from 5 to 40 km^3. This point underlines the advantage of using observed glacier volume changes to constrain model parameters, particularly in large basins with modest glacier coverage like Mica. Note that the second LHS gives higher maximum values of E because of the greater sampling density within the restricted parameter space and not necessarily because the glacier volume loss is close to the observed. More intense sampling within the parameter space that leads to higher glacier volume losses would likely lead to higher E at higher glacier volume losses as well. A substantial decrease in E can only be found with parameter combinations that lead to increases in glacier volume.

3.2 Model testing

Model testing on streamflow for the period 2000–2007, using the observed glacier extents from 2005, yielded an efficiency of 0.95 for the best model (Fig. 4a), a slightly better performance than during the 1985–1999 calibration period ($E = 0.93$). The goodness of fit of the best parameter set derived by the Latin hypercube search is essentially the same as the fit obtained by the combined evolutionary-steepest gradient optimization.

All 23 behavioral parameter sets reproduce the seasonal peak flows as well as low flows, but have difficulty with modelling intense rainfall events, especially during autumn (Fig. 4b). This is not surprising, since one of the two reservoirs (the slow reservoir) is primarily used to model the low flows during winter, and the single fast reservoir cannot simultaneously represent runoff generation due to melt and rainfall given the differences in their spatial patterns and nonlinearity. Since this model weakness only appears to affect rainfall-generated daily peak flows, it should not detract from the

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estimation of glacier melt contributions to streamflow, especially over monthly or longer time scales.

Despite the difference in spatial scales associated with modelled and observed SWE, SWE predicted by HBV-EC shows reasonable agreement with observations, with linear regressions between predicted and observed having R^2 of 0.82, 0.77, and 0.86 for the Molson Creek, Floe Lake, and Mount Revelstoke snowpillows, respectively (Fig. 5). The model accurately predicts the timing of the onset of snowmelt as well as the rate of decrease of SWE during the ablation period at all three snow pillow sites. However, the model tends to underestimate peak SWE. For some years this underprediction is within the expected error of SWE measurements (5 % according to Gray and Male, 1981). Snow pillows tend to overestimate SWE due to creep (downslope deformation of a snowpack), which puts additional load on the pillows (Gray and Male, 1981). However, there are some station-years in which the underestimation is too large to be simply attributed to measurement errors in the snow pillows (e.g., 1996–1997 at Floe Lake). The timing of these types of errors is not consistent among stations. For example, in the water year 1996–1997, peak SWE was reproduced reasonably accurately at Molson Creek and Mount Revelstoke, but strongly underpredicted at Floe Lake. Inconsistent variations in gauge catch efficiency could explain at least some of this underprediction. This inconsistent pattern of errors could also reflect, in part, the inherent variability in precipitation patterns from year to year, which are not properly represented through the use of fixed vertical gradients in each climate zone.

3.3 Historic contributions of glacier ice melt to streamflow

The mean annual contribution of ice melt to total streamflow varies between 3 and 9 % and averages 6 % (Fig. 6). Trend analysis revealed no significant increase or decrease of the annual contributions of glacier ice melt with time. For annual, August and September flows, the uncertainty bounds between runs with and without glaciers are close but do not overlap for most years (Figs. 6 to 8).

Contributions of glacier ice melt to discharge at Mica dam dominantly occur in August and September (Figs. 7 and 8). Mean August streamflow, calculated from the ensemble mean, would be up to 25 % lower if there were no glaciers, though the variation of contributions is relatively high (standard deviation = 7 % of the simulated mean flow with glaciers) (Fig. 7). The relative contribution of glaciers is highest in September, when ice melt can provide up to 35 % of the discharge. September contributions of ice melt are also less variable over time, with a standard deviation of 5 % of the simulated mean flow with glaciers (Fig. 8).

Figure 9 presents the mean and range of ensemble predictions for simulations with glaciers to simulations where glaciers have been removed and replaced by open land cover for two years with contrasting hydroclimatic conditions. Glacier runoff is particularly important in years with early snowmelt such as 1998, the year with the highest modelled ice melt (Fig. 6), where glaciers can contribute to more than 20 % of the flow for periods of more than two months. In years with late snow melt, such as the year 2000, glaciers have a minor effect on discharge. In July, some years have a higher discharge in the no-glacier scenario. This occurs because the glacier routing routine stores water in the early part of the melt season and releases it later. The glacier reservoir can lag flows from a few days up to several weeks, depending on the parameter values. This type of seasonal storage effect has been documented in previous studies of glacier hydrology (e.g., Stenborg, 1970).

4 Discussion

Unlike the findings reported by Stahl et al. (2008), the model efficiency does not clearly peak at the observed glacier volume loss, which is likely due to the lower glacier coverage in Mica and lower sensitivity of streamflow simulations to glacier parameters. Because Stahl et al. (2008) had winter mass balance measurements, they were able to fix the climate parameters, T_{lapse} and P_{lapse}, at an initial step during model calibration, separately from the calibration using streamflow data. This approach was not possible

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in this study as no mass balance measurements are available, so that a greater amount of uncertainty in these parameters (wider parameter ranges) is propagated through to streamflow predictions.

The guided GLUE approach for model calibration clearly demonstrates the value of glacier volume change for constraining model parameters that control snow accumulation and glacier mass balance. Given the general lack of mass balance data worldwide, our approach should prove useful for assisting in model calibration, particularly in large basins with modest glacier cover, where goodness-of-fit indices like the model efficiency are less sensitive to the streamflow variability related to glacier contributions. If a hydrologic model will be used to make future projections of the effects of climate change, it is imperative that a model be able to simulate glacier mass balance with reasonable accuracy, not just streamflow.

Given that repeat mapping may only be available over periods of a decade or more, the approach applied here will require relatively long calibration periods – 15 years in the case of Mica basin. Significant changes in glacier cover can occur over these periods, which should be accommodated during model calibration to avoid bias in the parameter estimates. For example, if too large a glacier area is used, the calibration may generate too small a melt coefficient to compensate, and vice versa. In this study, we split the calibration period into two sub-periods, each using a different glacier coverage. This approach necessitated the use of two five-year spin-up periods and rather complicated scripting using the R programming language to allow for an automated procedure. Even given the efficient structure of HBV-EC, which employs grouped response units and lumped reservoirs, the calibration process took substantial processing time (one week for 10 000 model runs on five CPUs, two weeks for the evolutionary optimization on one CPU). This challenge is not unique to the HBV-EC model, since most existing model codes with which we are aware do not allow for changes in land cover during a simulation run. One solution would be to develop a new model code that can accept updated land cover information without having to stop and restart execution.

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Glacier contributions to Mica basin streamflow were greatest in August and September. Although the contributions are relatively minor in terms of long-term average flows, they are most important during relatively warm, dry weather in summers following a winter with low snow accumulation and early snowpack depletion. These conditions can be critical from both water supply and ecological perspectives. Therefore, water managers and aquatic ecologists need to appreciate the hydrologic significance of glacier melt, even in large basins with moderate glacier cover.

5 Conclusions

Use of glacier volume change in the calibration procedure effectively reduced parameter uncertainty and helped to ensure that the model was accurately predicting glacier mass balance as well as the streamflow. This approach should be widely useful in glacier-fed catchments where mass balance observations are lacking. One drawback to the approach is that the calibration period must span the interval between glacier maps, in this case 15 years. Because glacier cover can change significantly over decadal and longer time periods – in this case it decreased by about 20 % – approaches are required to allow glacier cover to change through the calibration period.

While glaciers only cover 5 % of the Mica basin, they contribute up to 25 % of mean August flow and 35 % of mean September flow. These contributions are particularly important during periods of warm, dry weather following winters with low accumulation and early snowpack depletion. Glacier retreat over the twenty-first century could therefore have significant implications for streamflow during critical late-summer periods.

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Table 1. Description of calibrated model parameters, benchmark parameter values, parameter ranges for LHS, and correlation between each parameter and glacier net mass balance (b_n) and Nash-Sutcliffe efficiency (E).

Model routine	Parameter	Description	Benchmark Optimization	Parameter range Step 1		Correlation Step 1		Parameter range Step 2		Correlation Step 2		Parameter range Step 3		Correlation Step 3	
				from	to	b	E	from	to	b	E	from	to	b	E
Climate	TLAPSE	Temperature lapse rate (°C m^{-1})	0.00760	0.006	0.009	0.51	0.25	0.0075	0.009	0.48	−0.37	0.0075	0.009	0.58	−0.42
	PLAPSE	Fractional precipitation increase with elevation (m^{-1})	0.00003	0.00002	0.0005	0.22	−0.31	0.00002	0.0002	0.23	−0.10	5E-06	0.0001	0.28	−0.03
Snow	*AM*	Influence of aspect/slope on melt factor	0.12481	0	0.7	0.00	0.08	0.1	0.6	−0.01	−0.31	0	0.6	−0.11	−0.32
	CMIN	Melt factor for winter solstice in open areas (mm °C^{-1} day^{-1})	2.96739	2	4	−0.60	−0.07	2.2	3.4	−0.69	0.55	2.6	3.4	−0.50	0.33
	DC	Increase of melt factor between winter and summer solstice (mm °C^{-1} day^{-1})	0.18447	1	3	−0.44	−0.28	0.5	2	−0.42	−0.05	0	1.2	−0.45	0.10
	MRF-low	Ratio between melt factor in forest to melt factor in open areas below 1200 m	0.71696	0.4	0.9	0.00	0.00	0.6	0.8	0.00	−0.01	0.6	0.8	0.01	0.00
	MRF-high	Ratio between melt factor in forest to melt factor in open areas above 1200 m	0.72330	0.4	0.9	−0.02	0.00	0.6	0.8	−0.01	0.01	0.6	0.8	0.00	0.01
Glacier	MRG	Ratio of melt of glacier ice to melt of seasonal snow	1.04472	1	2	−0.33	−0.23	1	1.4	−0.24	−0.04	1	1.4	−0.33	0.00
	DKG	Difference between minimum and maximum outflow coefficients for glacier water storage (day^{-1})	0.02942	0.005	0.1		−0.07	0.01	0.05		0.01	0.01	0.05		0.01
Glacier (cont.)	KGMIN	Minimum outflow coefficient for glacier water (day^{-1})	0.01652	0.005	0.1		−0.01	0.01	0.05		0.09	0.005	0.03		0.09

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Table 1. Continued.

Model routine	Parameter	Description	Benchmark Optimization	Parameter range Step 1		Correlation Step 1		Parameter range Step 2		Correlation Step 2		Parameter range Step 3		Correlation Step 3	
				from	to	b	E	from	to	b	E	from	to	b	E
Runoff	KF	Proportion of fast reserviour release (day^{-1})	0.14108	0.01	0.4		0.23	0.1	0.35		−0.19	0.05	0.3		0.18
	ALPHA	Exponent to adjust release rate of fast reservoir	0.13276	0.01	0.3		0.04	0.05	0.2		−0.11	0.05	0.2		0.05
	KS	Proportion of slow reservoir release (day^{-1})	0.01442	0.0005	0.05		−0.04	0.001	0.03		0.15	0.0005	0.015		0.07
	FRAC	Fraction of runoff directed to fast reservoir	0.89401	0.5	0.9		0.11	0.7	0.9		0.35	0.7	0.9		0.49

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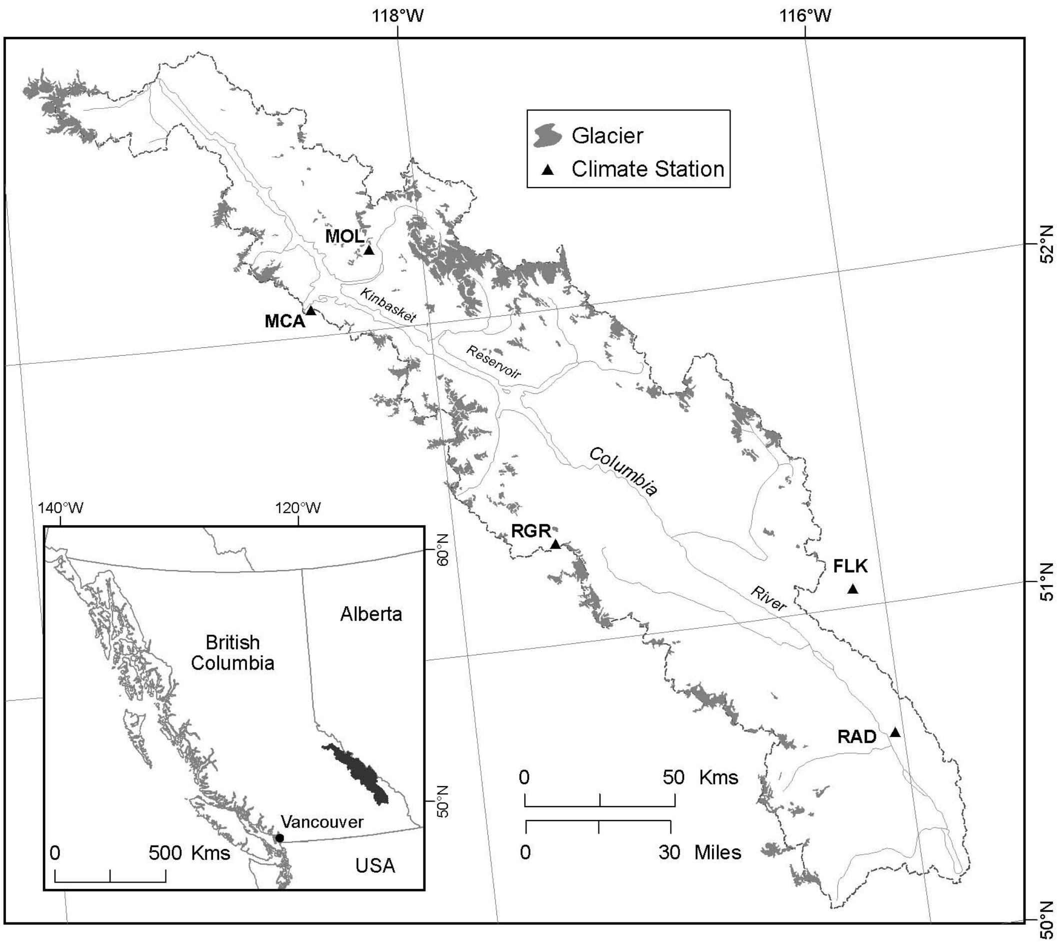

Fig. 1. Location of the Mica basin and locations of climate stations used to force the hydrological model.

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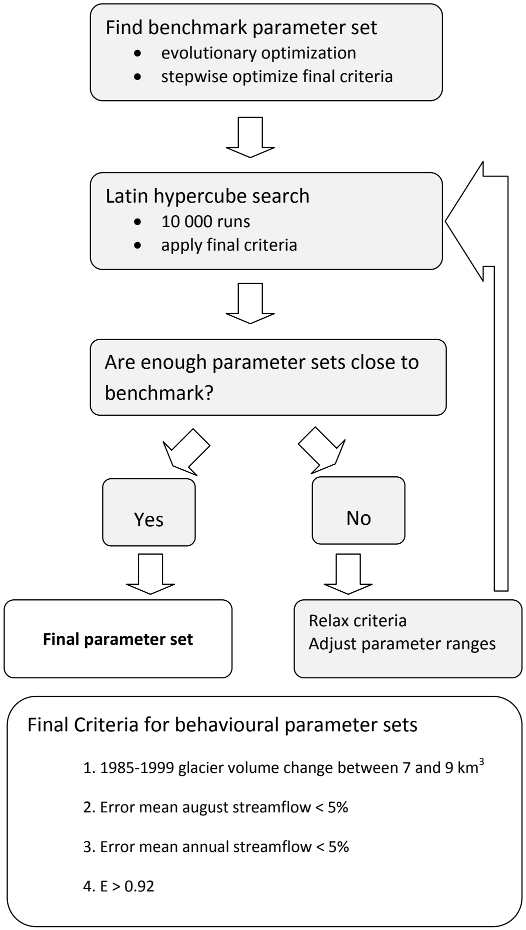

Fig. 2. Flow chart illustrating the "guided" GLUE approach for calibration and uncertainty analysis. E is the Nash-Sutcliffe efficiency.

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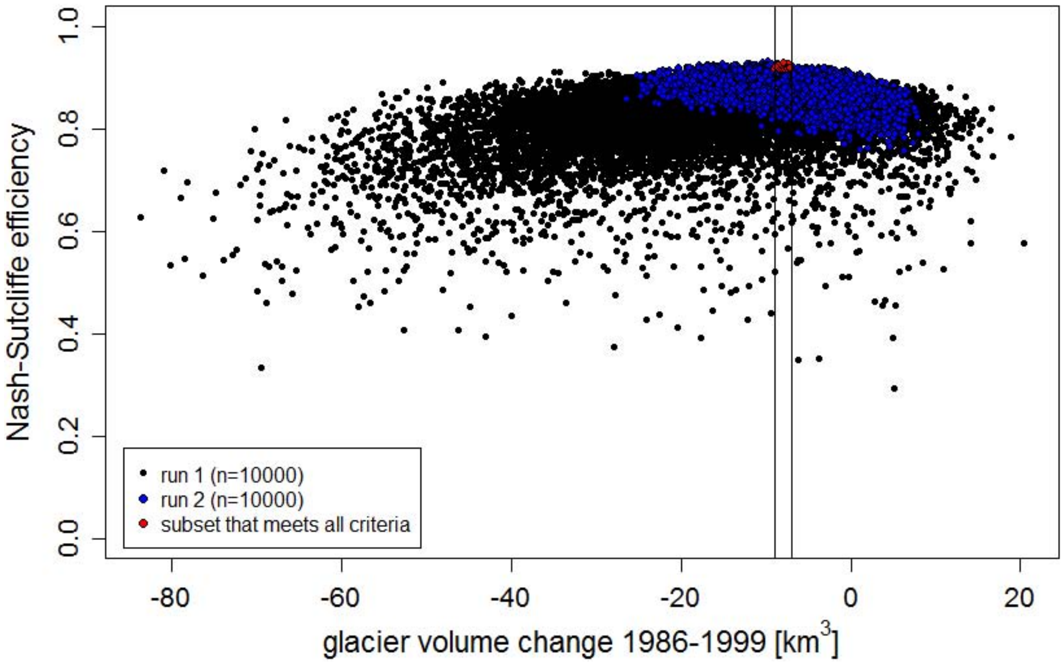

Fig. 3. Nash-Sutcliffe efficiency (E) plotted against simulated glacier volume change for 10 000 model runs in the initial Latin Hypercube Search (black) and for 10 000 model runs in a Latin Hypercube Search with adjusted prior parameter distributions (blue). Red dots indicate acceptable parameter combinations.

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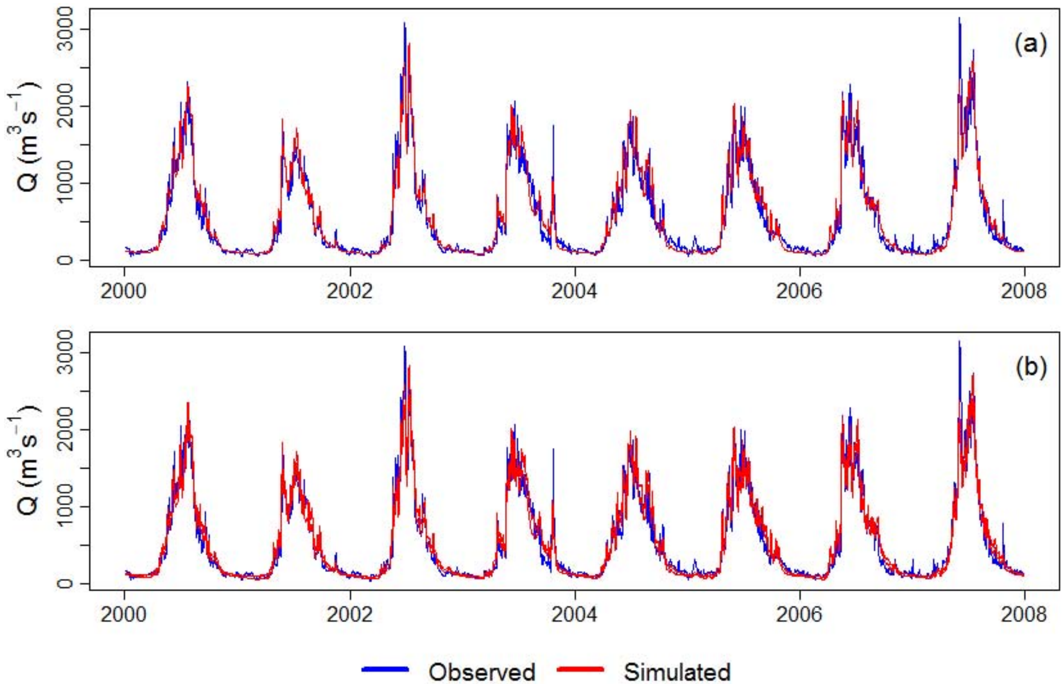

Fig. 4. Observed and simulated discharge for the test period (2000–2007). **(a)** Observed and simulated discharge predicted with the best performing (E) parameter set; **(b)** observed and the ensemble of simulated discharge.

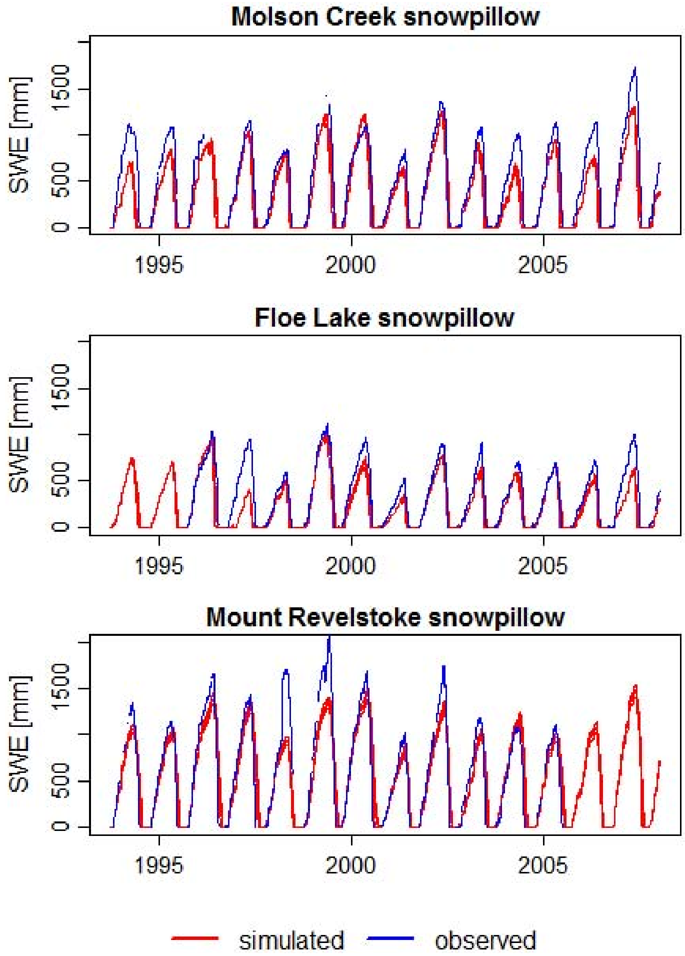

Fig. 5. Simulated (using the best model) and observed snow water equivalents for three snow pillow sites. Simulations are for the GRU that corresponded to the snow pillow sites.

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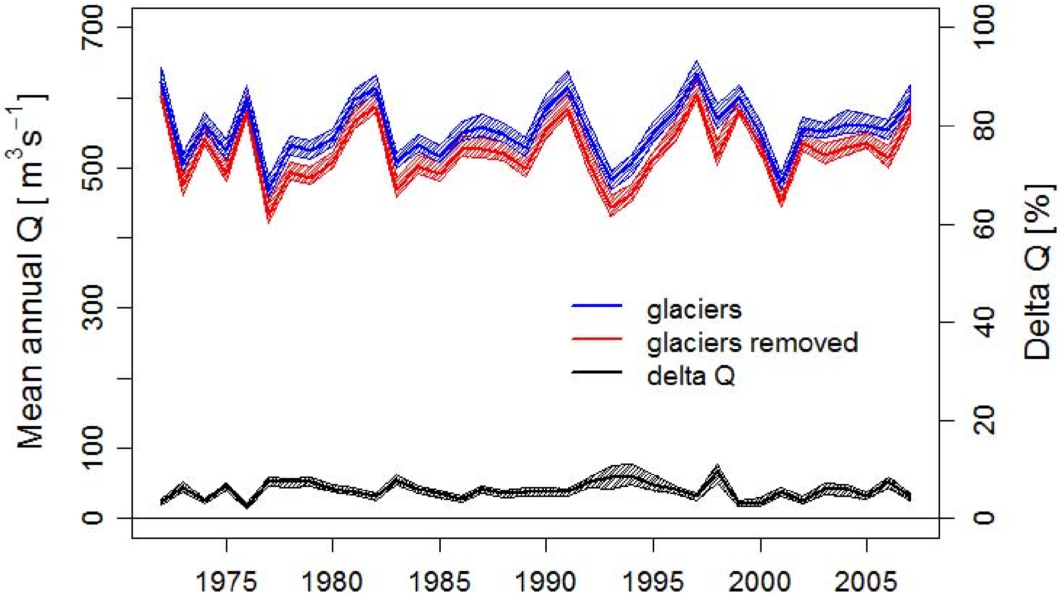

Fig. 6. The effect of glaciers on mean annual discharge shown by comparing simulations with and without glaciers including uncertainty limits (5–95 % quantile range).

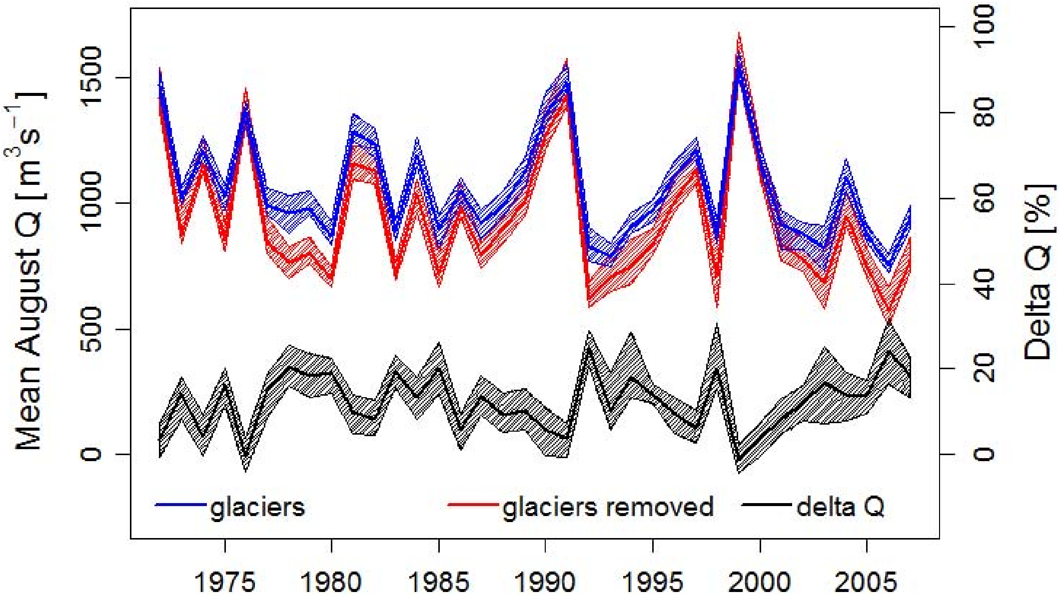

Fig. 7. The effect of glaciers on mean August discharge shown by comparing simulations with and without glaciers including uncertainty limits (5–95 % quantile range).

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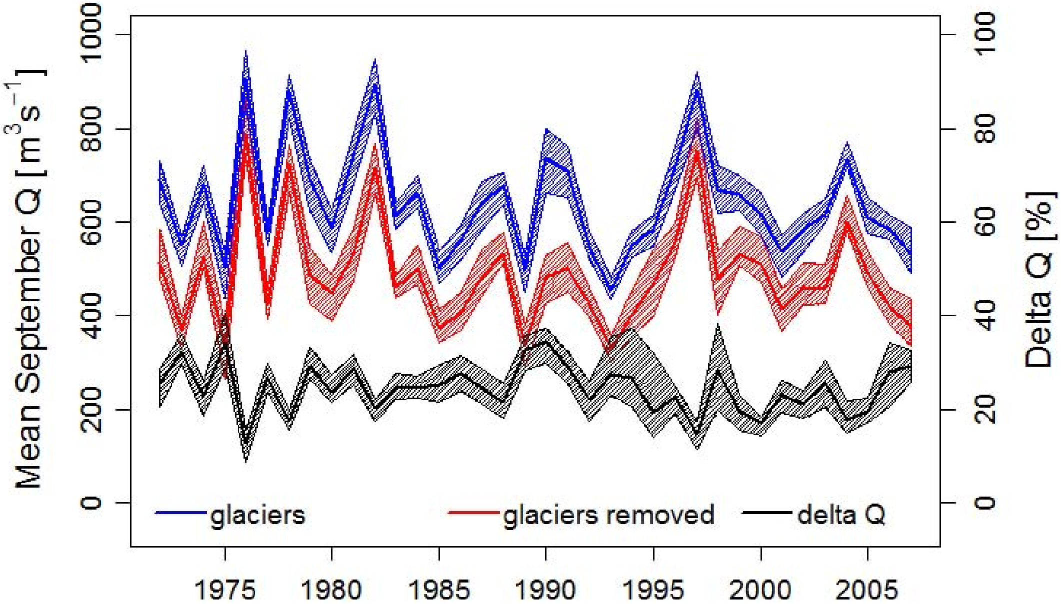

Fig. 8. The effect of glaciers on mean September discharge shown by comparing simulations with and without glaciers including uncertainty limits (5–95 % quantile range).

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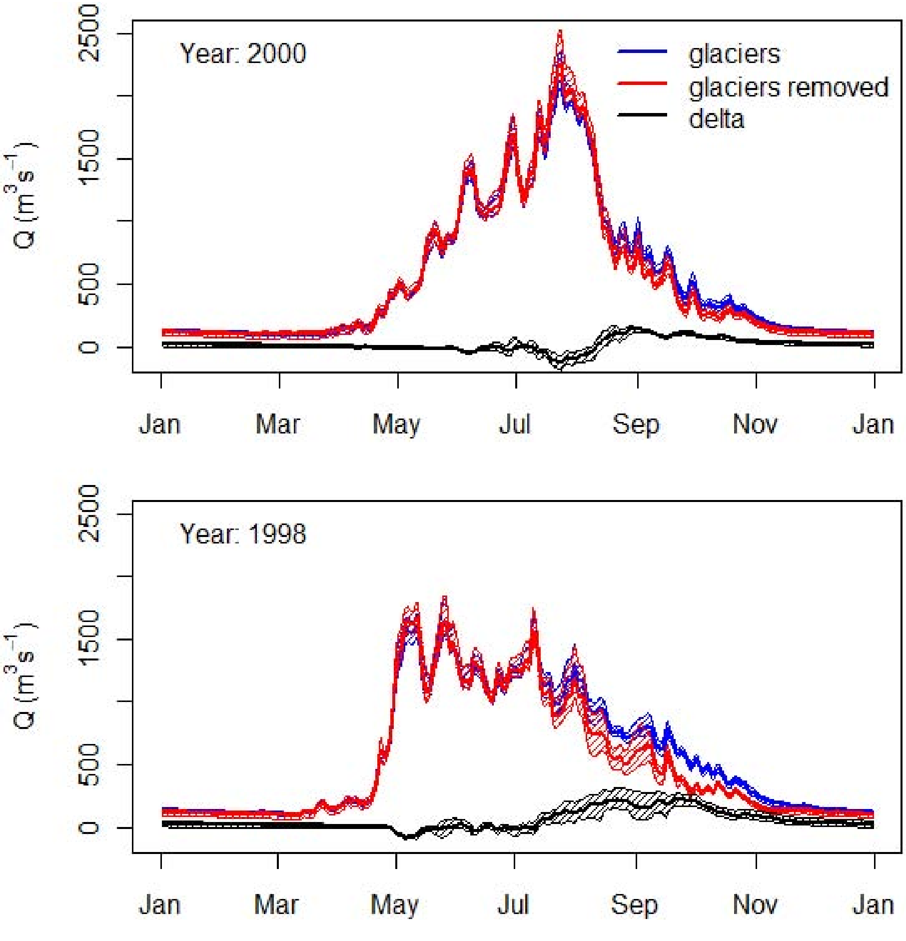

Fig. 9. The effect of glaciers on discharge shown by comparing simulations with and without glaciers for the year with the highest modelled ice melt (1998) and the year with the lowest ice melt (2000).

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