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Peak river flows in cold regions – Drivers and modelling using GRACE satellite observations and temperature data

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Running title: Forecasting peak river flows in cold regions

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Abstract

The peak river flow for the Mackenzie River is modelled using GRACE satellite observations and temperature data, which advances the applications of space-based time-variable gravity measurements in cold region flood forecasting. The model estimates peak river flow by simulating peak surface runoff from snowmelt and the corresponding baseflow. The modelled results compared fairly well with the observed values at a downstream hydrometric station. The results also revealed an average 22-day travel time for the snowmelt water to reach the hydrometric station. The major driver for determining the peak flow was found to be the temperature variations. Compared with the Red River basin, the results showed that the Mackenzie River basin has relatively high water storage and water discharge capability, and low snowmelt efficiency per unit temperature. The study also provides a GRACE-based approach for basin-scale snowfall estimation, which is independent of in situ measurements and largely eliminates the limitations and uncertainties with traditional approaches. The model is relatively simple and only needs GRACE and temperature observations for peak flow or flood forecasting. The model can be readily applied to other cold region basins, and could be particularly useful for regions with minimal data.

Key Points:

- Peak runoff and winter baseflow are modelled using GRACE observations.
- Major drivers controlling peak river flows or floods in cold regions are identified.
- Mechanisms underlying peak flow regimes of different basins are discussed.
- GRACE observation is used for basin-scale snowfall estimate.

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1. Introduction

Peak river flow and its forecasting are of considerable interest to flood control and emergency service agencies, river recreationists, wildlife managers, hydropower plant operators and anyone interested in the combined effect of watershed yield and human regulation on the river maximum flows at a site. Flooding is a hazard to developed areas and to human activities in the floodplain and causes human hardship and economic loss (International Joint Commission. 2000; Burton et al., 2016). Flooding ranks as one of the most damaging forms of natural disasters in the world. On the other hand, flooding is essential to a healthy environment. Floods can benefit the natural environment and sustain many ecosystems. For instance, the recurrent inundation of the Peace-Athabasca Delta in the Mackenzie River basin in Canada has fostered an environment in which plant and animal life has achieved a balance dependent on flooding. In fact, water spilling across the floodplain nourishes the wetlands of many of Canada's major deltas (Andrews, 1993).

Peak river flow in cold regions is commonly a result of snowmelt during the spring break-up, or the subarctic nival regime (Church, 1974). During the winter most of the precipitation is in the form of snow, which accumulates in the basins until the spring breakup. Depending on the temperature in the spring freshet, huge quantities of water from snowmelt can be produced. Meltwater is unable to penetrate when the ground is frozen and runs off over the ground surface into rivers and lakes, resulting in peak flows and floods. Snowmelt runoff floods are the most common type of flooding in Canada (Andrews, 1993; Yang et al., 2009; Rannie, 2015; Wang and Russell, 2016).

Two key factors determining the magnitude of peak river flows or floods during the freshet are the amount of snow water equivalent (SWE) available at the time of spring breakup

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and the temporal patterns of rising air temperature during the snowmelt season. As such, forecasting peak flows or floods for rivers in cold regions, regardless of the methods being used, heavily relies on information for basin-scale SWE and temperature. SWE can be estimated by the difference of snowfall and water loss from snow sublimation. Unfortunately, accurately measuring snowfall is difficult and often involves large uncertainties. Studies show that the measurement errors of snowfall at climate stations can be as high as 50% due to the wind-induced undercatch and the fact that many trace events cannot be easily measured (Goodison et al., 1998). Snow sublimation estimation is also difficult and has large uncertainties (Wang et al., 2013; Wang et al., 2015a). In addition, cold regions commonly have very sparse observational data. For instance, the density of climate stations in the Canadian Northwest Territory is only about 2 stations per 100,000 km^2 (Metcalfe, 1994). This sparse spatial density of stations makes the up-scaling of SWE from site to basin-scale extremely difficult and unreliable. Remote sensing, either by optical (e.g., Hall et al., 2002) or microwave (e.g., Dong et al., 2005) satellite sensors, can provide valuable information for the spatial snow cover distribution. However, remote sensing approaches for the estimation of SWE depend on retrieving snow parameters such as snow cover area, snowpack depth, density, liquid and ice content, etc., which can be significantly affected by the climate/atmosphere and land surface/snowpack conditions, such as the temperature effect on snow density, and cloud, vegetation cover and topographic impacts on snow cover area identification (Nolin, 2010). Moreover, SWE models from remote sensing heavily rely on in situ data for calibration and validation, which may propagate the errors in the in situ measurements to the remote sensing products (Hall and Riggs, 2007). Due to the difficulties mentioned above, spatial snow data products either from traditional methods or from remote sensing approaches often have large uncertainties. For instance, recent studies have found

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that errors in snow data are the principal contribution to the water budget imbalance in cold-region basins (Wang et al., 2014a; Wang et al., 2014b; Wang et al., 2015b). Consequently, accurate estimation of basin-scale SWE is a key challenge in improving flood forecasting over cold region rivers.

In a recent study, Wang and Russell (2016) developed a flood forecasting model using the Gravity Recovery and Climate Experiment (GRACE) satellite observations and temperature data. In the model, the total water storage (TWS) data from GRACE was used to estimate the basin-scale SWE at the spring breakup. The model does not require snowfall and sublimation data as well as the site-to-basin upscaling process, thus largely eliminating the data constrains discussed above. In Wang and Russell (2016), the model was applied to the Red River basin, a USA-Canada transboundary watershed located in central North America, mostly in the U.S. states of Minnesota and North Dakota. Floods predicted by the model compared well with the observed values. However, the Red River has a drainage area of 116,500 km^2, which is limited in size relative to the large footprint of the GRACE satellites. This size limitation may contribute to the large data uncertainties in the TWS of the basin which might substantially impact the accuracy of flood forecasts. Second, a conclusion found in Wang and Russell (2016) is that the SWE at the spring breakup is the major driver determining the magnitudes of peak river flows, with temperature playing the secondary role. Whether this conclusion is valid for other basins with different hydroclimatic conditions still remains to be investigated.

In this study, the model of Wang and Russell (2016) is calibrated to forecast the peak flows for the Mackenzie River in Canada (Fig. 1). The Mackenzie River basin has a large drainage area of 1.8 million square kilometres, which largely eliminates the constraints imposed by the large footprint of GRACE. The basin is located in northwest Canada, about 18 degrees of

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latitude north of the Red River basin. Compared with the Red River basin, the Mackenzie River basin has very different physiographic and hydroclimatic conditions, including the very low temperature, large amount but small interannual variations in winter snowfall, as well as high water storage conditions. As a result, the Mackenzie River has much delayed peak discharge flows that have much smaller interannual variations than the Red River. The objective of this study is to evaluate the model performance in forecasting, and to identify the major drivers in determining, the peak flows or floods for the Mackenzie River. The model results and parameter values will be compared with that for the Red River to help better understand the mechanisms underlying the peak river flows for basins with different physiographic and hydroclimatic conditions.

2. Study Region, Data, and Hydroclimate Characterization

The Mackenzie River Basin is one of the largest North American river basins and covers about 1.8 million square kilometres or about 20% of the landmass of Canada. Our basin study area is located between 52° to 70° N and encompasses three major physiographical regions. In the west, the Western Cordillera consists of a series of mountain chains and valleys or high plateaus. Many ridges of the Rocky Mountain chain exceed 2000 m elevation, and some have glaciers occupying the mountain tops and high valleys. To the east is the Canadian Shield, a rolling terrain with myriad of lakes and valley-wetlands separating upland outcrops of Precambrian bedrock. The central zone is part of the Interior Plains, with wetlands, lakes, and vegetation that ranges from prairie grassland in the south, through the boreal and subarctic forests, to the tundra in the north (Woo and Thorne, 2003). There are a large number of lakes across the basin, with the top three large lakes (i.e., Great Bear Lake, Great Slave Lake and Lake Athabasca) having a total surface area of over 67,000 km^2. The basin has a wetland coverage of approximately 49% and permafrost underlies approximately 75%. The Mackenzie River system flows a total of 4,241 kilometres from its headwaters northwards to the Arctic Ocean.

The data required by the model for peak flow estimation includes GRACE TWS and daily temperature T_a. The in situ measurement of river flow Q is used for model calibration. The GRACE Release-05 TWS datasets were downloaded from the GRACE Tellus website (ftp://podaac-ftp.jpl.nasa.gov/allData/tellus/L3/land_mass/RL05/) (Swenson. 2012). The datasets include monthly TWS from three data processing centers: CSR (University of Texas/Center for Space Research), GFZ (GeoForschungsZentrum Potsdam), and JPL (Jet Propulsion Laboratory). The data are provided in a spatial sampling of 1 degree grids in both latitude and longitude. The land grid scale factor, as provided with the TWS data, was applied to recover signal loss due to filtering and truncation (Swenson and Wahr, 2006). Monthly error estimates in the GRACE TWS for the Mackenzie River Basin are based on combined measurement and leakage error, following Wahr et al. (2006). A detailed description of the data processing and accuracy assessment can be found in Landerer and Swenson (2012). The differences among the three datasets, as shown later, were found to be small over our study region. In this study, the average TWS of the three datasets was used. The study covers 12 snow-years from the fall of 2002 to the spring of 2014. The baseline of the TWS, which was based on the average from January 2004 through December 2009 in the original data, was re-adjusted to the minimum value found over the 12-year period.

The basin average T_a was calculated from the Global Land Data Assimilation System (GLDAS) meteorological forcing of temperature, which is provided at a 3 hour time step and a spatial resolution of 0.25×0.25 degree latitude/longitude. The air temperature time series for 2002-2014 was downloaded from Goddard Earth Sciences Data and Information Service Center (http://mirador.gsfc.nasa.gov/). The daily air temperature of the grids is taken as the average of the 8 readings of the day, and the daily T_a of the basin is the average of the daily values of all the GLDAS grids within the basin. The GLDAS precipitation is also processed to assist our discussion. The daily precipitation of each grid is the sum of the 8 readings of the day, and the daily value of the basin is the average of all the grids within the basin.

The in situ daily Q observed at the hydrometric station of Mackenzie River at Arctic Red River (station ID: 10LC014, Latitude: 67° 27' 21" N, Longitude: 133° 45' 11" W) was obtained from the Water Survey of Canada (http://www.ec.gc.ca/rhc-wsc/). The Q collected at this location, before the river branches into many distributaries, is considered as the total flow for the Mackenzie River system. The station controls an area of 1,679,100 km² (> 90% of the basin). The original Q is in $m^3\ s^{-1}$ and it was converted to water depth (mm) using the basin area. The peak flow is the maximum mean daily flow (the highest average flow for an entire day) in a year at the station. It is worth noting that the Q data is only required for model calibration. The peak flow estimation, once the model is calibrated, will only need TWS and T_a prior to winter and at the snowmelt season in spring.

The Mackenzie River Basin has several climatic regions, including cold temperate, mountain, subarctic, and arctic zones. According to the climate data used in this study, the daily T_a averaged for the basin ranged from a low of -36 ºC to a high of 21 ºC during the 12 study years (Fig. 2). The 12-year average T_a of the basin varied from -22.6 ºC in winter to 16.4 ºC in summer, with an annual mean of -2.4 ºC. The temperature dropped below 0 ºC in mid-October and rises above 0 ºC at the end of April, with an annual average of more than 200 days having temperatures below 0 ºC. In winter, the basin is in a deep frozen state and precipitation as snowfall accumulates until spring breakup. Annual precipitation over the 12 years ranged from a low of 311 mm to a high of 432 mm, with an average of 386 mm (Fig. 2). Annual precipitation during the winter time when T_a<0 ºC ranged from 124 mm to 189 mm, of which the average was 157 mm, or 41% of the total annual precipitation. However, it is generally recognized that snowfall from the climate station measurements is underestimated. This will be further discussed using our GRACE-based estimates in Section 4.

The TWS of the basin has strong seasonal variations, with highest values around April before snow breakup and lowest values around October before snow starts to accumulate (Fig. 2). The average seasonal variation range (peak to peak) for the 12 years was 116 mm. The pronounced seasonal pattern is mainly a result of snow accumulation combined with extremely low evapotranspiration and river discharge in winter, and high evapotranspiration combined with high river discharge in summer (Wang and Li, 2016). The TWS has moderate interannual variations. Among the 12 years, the maximum differences reached 68 mm in April and 77 mm in October. The three original TWS datasets have an average difference of 5.4 mm, which is minor compared with the seasonal variations of TWS for the basin. The combined measurement and leakage error for the GRACE TWS datasets is estimated at 9.9 mm, which is about 8.5 % of the average seasonal variation of TWS.

River flow of the Mackenzie River is greatest in spring when snowmelt occurs (Fig. 2). The peak flows for the study period observed at the station 10LC014 varied from a low of 24,200 $m^3 s^{-1}$ in 2010 to a high of 35,000 $m^3 s^{-1}$ in 2013 (Fig. 2). The 12-year average peak flow was 28,400 $m^3 s^{-1}$. The date when peak river flow occurred was on June 3 on average, with the earliest on May 15 in 2005 and latest on July 12 in 2007. In the climate change context, studies have showed that the Mackenzie River is experiencing a shift of peak flows to earlier in the spring due to earlier melt of snow cover and river ice breakup (Prowse et al., 2010; Woo and

Thorne, 2003; Yang et al., 2014). The basin discharge in winter gradually decreases and at the time of spring breakup, the observed 12-year average flow was around 3,700 m^3 s^{-1}. In summer, although the total amount of rain was higher than the winter snowfall, the river discharge peaks were significantly lower (Fig. 2), mainly due to the high evapotranspiration, recharge of water to the soil and aquifers, and relatively even distribution of rain over the season. The large lakes in the basin also provide natural regulation to the system.

3. Methods

3.1. Model Description

The model includes five major steps that are summarised below.

Step 1: Determining snow season and the total water storage change

This study only concerns the snow season, defined as the time period from the start (t_0) of snow accumulation in late autumn to the snow breakup (t_b) in the next spring (Fig. 3). The t_0 is determined by the criteria of (1) the first precipitation event in late fall when daily average temperature T_a drops below 0°C and (2) the accumulated T_a after this date remains negative. The criteria are to ensure that precipitation is in the form of snow and to exclude temporary snowfalls that are likely to melt in early winter. After this date, the net snow amount, i.e., snowfall minus snow sublimation, is accumulated over the basin in winter during which the T_a remains far below 0°C (Fig. 2). The spring breakup time t_b is determined by the criteria of (1) daily average temperature T_a rises above 0°C and (2) the accumulated T_a after this date remains positive. The criteria are to ensure that the heat conditions of the basin will lead to the spring breakup and to exclude the possible minor snowmelt events in the snow cover season in case the T_a momentarily rises above 0°C.

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The basin total water storages at t_0 and t_b, i.e., TWS_0 and TWS_b, are estimated by linearly interpolating the GRACE TWS for the two months before and after t_0 and t_b, respectively. The TWS_0 represents the maximum amount of non-snow water in the basin during the snow cover season. It mainly consists of surface water, soil water, and groundwater, part of which will be discharged in winter. The TWS_b represents the sum of net snow accumulation (S_b) and non-snow water content left in the basin at time t_b (Fig. 3).

Step 2: Modelling baseflow

In winter due to the frozen soil and snow cover, water infiltration and evaporation at the soil surface is minimal. The decrease of non-snow water (W) in the basin, $dW(t)/dt$, is thus mainly due to basin discharge or baseflow. In this study, the winter baseflow Q_{base} is modelled using a first order differential equation:

$$Q_{base}(t) = -dW(t)/dt = a(W(t) - b) \quad (1)$$

where a is a parameter representing the lump conductivity of the basin for water discharge, and b is a parameter representing the threshold value of W at which the basin would have zero discharge. This simplified model represents the basin water discharge as proportional to the available water storage and lump basin conductivity for discharge.

With the above model and the initial condition of $W(t_0)=TWS_0$, the accumulated total baseflow in winter, Q_{sum}, can be determined by:

$$Q_{sum} = \int_{t_0}^{t_b} Q_{base}(t)dt = TWS_0 - (TWS_0 - b)e^{-a(t_b - t_0)} - b \quad (2)$$

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The values for parameters *a* and *b* are obtained numerically by finding the least square errors between the observed and modelled Q_{sum} for the 12 study years. Once *a* and *b* are known, the baseflow rate at any time from t_0 can be calculated as:

$$Q_{base}(t) = a(TWS_0 - b)e^{-at}. \tag{3}$$

Step 3: Determining snow water equivalent at spring break-up time

After Q_{sum} (Eq. 2) is known, the snow water equivalent at t_b, S_b, can then be obtained as the sum of Q_{sum} and the change of TWS in the snow season (i.e., $TWS_b - TWS_0$) based on water balance of the basin (Fig. 3):

$$S_b = TWS_b - TWS_0 + Q_{sum} \tag{4}$$

The S_b represents the initial amount of snow which is going to melt following the temporal trajectory of air temperature as described next. Note that snow sublimation in winter is implicitly included in the estimate of S_b as GRACE measures the net change of water storage, which is contributed by the difference of snowfall and snow sublimation in winter.

Step 4: Modelling snowmelt (*M*) and peak surface runoff (Q_{runoff})

The snowmelt is estimated by a temperature index model at a daily time step:

$$M(t) = \min(S(t), \alpha(T_a(t) - \beta)) \tag{5}$$

where *M(t)* and *S(t)* is the snowmelt and snow amount available on day *t* (Note *S(t)=S(t-1)-M(t-1)*), respectively, β is a base temperature for snowmelt and α is the snowmelt rate per unit of temperature above β. Equation (5) determines the actual snowmelt rate by both temperature and snow availability on a given date. The time series of snowmelt is calculated using the initial

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condition of S_b (Fig. 3) and the daily T_a time series. The parameter values for α and β were solved using a nested numerical iteration scheme to find the best correlation between peak snowmelt rate and the observed peak surface runoff. Observed peak surface runoff is calculated as the difference between observed peak river flow and the corresponding baseflow obtained using Equation (3). The model for estimating peak surface runoff, Q_{runoff}, from peak snowmelt is obtained after this numerical process is done.

Step 5: Modelling peak river flow

The modelled peak river flow is the sum of peak surface runoff and the corresponding baseflow obtained above:

$$Q_{peak} = Q_{base} + Q_{runoff} \tag{6}$$

In summary, the model determines peak river flows by simulating its two components of peak surface runoff from snowmelt and the corresponding baseflow from basin discharge. The snowmelt is simulated by a temperature index model. The basin discharge is simulated by a first order differential equation model. There is a total of four parameters, a and b, which are calibrated using the total winter flow, and α and β, which are calibrated using the observed peak surface runoff. Once calibrated, the model only needs GRACE TWS and T_a as inputs for determining peak river flow after spring breakup.

3.2. Model Evaluation

The model results were compared to in situ Q observations and evaluated using mean absolute error (MAE), the Pearson correlation coefficient (r) and t-test for significance levels (p), as well as the Nash–Sutcliffe model efficiency coefficient (E). The E is commonly used to assess

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the predictive power of hydrological models. The leave-one-out cross-validation (LOO-CV) approach is used to evaluate the general model performance in peak river flow forecasting. The result from this LOO-CV evaluation is generally a more conservative estimate of the model performance than that trained on all samples. The model results were also compared with other basins to understand the mechanisms underlying the variations of peak flows over different regions.

4. Results and Discussion

The model gives an average estimate for the start of snow cover of t_0 on October 14 and spring breakup of t_b on April 29. The average snow water equivalent accumulated during this time period (197 days) is S_b=160 mm. Given a water loss of 22 mm due to snow sublimation based on Wang et al. (2015a), the total snowfall during this time period from t_0 to t_b is thus 182 mm. In comparison, the corresponding total precipitation amount from the GLDAS datasets as discussed in Section 2 is 150 mm (snowfall=133 mm, rain=17 mm), which is about 20% lower than the GRACE-based estimate. Underestimation in snowfall over cold regions has been recognised by many studies (e.g., Wang et al., 2014a; Wang et al., 2015b) due to the errors as discussed in the Introduction section. This is particularly so for our study region as the available benchmark measurements, while extremely sparse, are mostly located in valleys or lowlands where the data are likely to underestimate the average condition of the entire basin. Many efforts have been made to correct the biases (e.g., Woo et al., 1983; Yang et al., 2005) but they are largely constrained by the difficulties in quantifying snow amounts at basin-scale. Our study provides a new GRACE-based approach for estimating snowfall at basin-scale. The approach is largely independent of in situ snowfall data thus eliminates most of the limitations and uncertainties in the snow gauge measurements and the site-to-basin up-scaling processes.

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The model results show that the basin has a lump conductivity of a=1.07×10^{-3} day^{-1} for water discharge, and a threshold value of b=-195.9 mm of water below which the basin would have no discharge (Table 1). Compared with the Red River basin (Wang and Russell, 2016) which has a=0.49×10^{-3} day^{-1} and b=9.2 mm, the model results suggest that the Mackenzie River basin has a relatively high basin conductivity and high water storage state. This is consistent with the facts that the Mackenzie River basin has about 49% of its area as wetland and a large number of lakes which are highly effective in providing large storage capacities and are available for water discharge. Annual evapotranspiration for the basin is less than 60% of its annual precipitation (Wang et al., 2015a), so the basin has a large water surplus for soil and aquifer recharge as well as lake and wetland replenishment to sustain the winter low flows. In contrast, the Red River basin has a very flat terrain (the slope of the river averages less than 10 cm per kilometer). It also has high evapotranspiration that is close to its precipitation in summer, resulting in a very low water storage state prior to winter. The model results are also agreeable with the observed difference in winter flows, which are as low as 5.8 mm day^{-1} for the Red River and as high as 48.8 mm day^{-1} for the Mackenzie River as discussed next.

The model performance for winter baseflow estimation is shown in Table 1 and Fig. 4A. Compared to the mean observed baseflow of 48.8 mm in winter over the study period, the model has a mean absolute error of MAE=2.37 mm, or 4.9% of the observed value. The modelled winter baseflow for the 12 years has a correlation coefficient of r=0.73 with the observed values at a significance level of p<0.007. The Nash–Sutcliffe model efficiency coefficient for baseflow estimates is E=0.53. The results suggest that the basin winter discharge is primarily driven by its water storage prior to winter. Indeed, surface runoff is minimal in winter due to the lack of liquid precipitation. Water exchange between soil and groundwater is also minimal due to the frozen

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soil. River flow in winter is thus sustained by groundwater and lake discharge that is controlled by pre-winter storage conditions. Determining groundwater and lake water storage at the basin scale is extremely difficult by traditional methods. The above results underscore the advantages of using GRACE satellite observations in estimating basin water storages and discharge in winter.

The snowmelt model suggests a snowmelt rate of α=17.0 mm per unit of temperature above a base temperature of β=2.1°C (Table 2) for the Mackenzie River basin. The results are close to that obtained for the Red River basin which has α=18.2 mm °C^{-1} and β=1.0°C (Wang and Russell, 2016). The slightly higher melting power for a unit temperature with a lower base temperature for the Red River basin reflects the impacts of other environmental variables on the snowmelt, such as higher solar radiation for the Red River basin than for the Mackenzie River basin. Comprehensive and physically based snowmelt models are available and they have the advantages of simulating the integrated impact of all environmental variables (e.g., radiation, humidity, wind speed) on the snowmelt processes (e.g., Wang et al., 2007; Wang, 2008; Zhang et al., 2008), but these kinds of models are data demanding and difficult to implement for operational use over data scarce regions. We use the temperature index model as it needs minimal data input and is computationally simple. Our results also show that the temperature index model performs fairly well, consistent with many other studies (e.g., Li and Simonovic, 2002; Griessinger et al., 2016).

The model performance for estimating peak surface runoff Q_{runoff} is shown in Table 2 and Fig. 4B. Compared with the observed mean Q_{runoff} of 1.26 mm day^{-1} over the study period, the model has a mean absolute error of MAE=0.1 mm day^{-1}, or 7.6% of the observed value. The modelled Q_{runoff} for the 12 years has a correlation coefficient of r=0.82 with the observed values

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at a significance level of p<0.001. The Nash–Sutcliffe model efficiency coefficient for surface runoff estimates is E=0.50, slightly lower than that for the baseflow estimates.

To further determine the relative importance of T_a and S_b in Q_{runoff}, we analysed the relationship between S_b and Q_{runoff} (Fig. 5) and found that Q_{runoff} showed little correlation with S_b. The result indicates that total snow amount at spring breakup has little impact on the Q_{runoff}. As such, the rising temperature during the snowmelt season is the main driver for determining the Q_{runoff} for the Mackenzie River basin. In contrast, the correlation between S_b and Q_{runoff} for the Red River basin was found to be fairly strong (Fig. 5). Specifically, without including T_a, the S_b by itself explained more than two thirds (the coefficient of determination r^2=0.673) of the interannual variations in Q_{runoff}, suggesting that the major driver for Q_{runoff} is S_b for the Red River basin (Wang and Russell, 2016). In another study by B.C. Ministry of Forests, Lands and Natural Resource Operations (2012), the peak flows for the Lower Fraser River were analysed. The Lower Fraser River is located in the latitudes between the Mackenzie River basin and the Red River basin. It was reported that the snow factor contributes about 20-40%, and the weather factors (mainly temperature) contribute about 60-80% to the flood risk. The above results for the three basins are consistent and appear to suggest that the principal drivers for peak river flows vary between basins. For northern basins temperature plays an important role, whereas for more southern basins the amount of snow accumulation is more important.

The difference in the main drivers for Q_{runoff} for the basins is largely due to the difference in their hydroclimatic conditions. The Red River basin has a mean snow accumulation at spring breakup of S_b=73.3 mm, which is less than half the accumulation for the Mackenzie River basin (S_b=160.0 mm). On the other hand, the Red River basin has a much larger interannual variations of S_b than that for the Mackenzie River basin. The coefficient of variation (CV) of S_b, or relative

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standard deviation (RSD), which is calculated as the ratio of one standard deviation to the mean, is as high as 49.2% for the Red River basin. In contrast, the CV is only 14.7% for the Mackenzie River basin. The small S_b with its large interannual variations lead to the fact that years with large S_b often correspond to severe floods, and years with small S_b often correspond to very low flows, in the Red River (Wang and Russell, 2016). For the Mackenzie River basin, large and relatively stable snow amounts lead to the fact that the interannual variations of Q_{runoff} are very small and they are mainly determined by the temperature anomalies during the snowmelt season. The identification of major drivers for peak surface runoff for different basins is of importance in river flow modelling and flood forecasting.

The model performance for estimating peak river flow Q_{peak}, based on the above results for Q_{base} and Q_{runoff}, is shown in Table 3 and Fig. 4C. Compared to the mean observed Q_{peak} of 1.46 mm day^{-1} (28,400 km^3 s^{-1}) over the study period, the model result has a mean absolute error of MAE=0.1 mm day^{-1} (1,878 km^3 s^{-1}), or 6.5% of the mean Q_{peak} value. The modelled Q_{peak} for the 12 years has a correlation coefficient of r=0.83 with the observed values at a significance level of p<0.001. The Nash–Sutcliffe model efficiency coefficient for peak river flow estimates is E=0.51. Of the peak river flow, 15% is contributed by baseflow and 85% by surface runoff. As such, the modelling accuracy in Q_{base} plays a small role in the modelling accuracy of peak river flows or flood forecasts. However, modelling accuracy for total winter baseflow (Q_{sum}) could be of importance as Q_{sum} directly affects the estimate of S_b at spring break-up (Fig. 3), which is the case for the Red River basin. Compared to the dates for peak snowmelt, the dates for the peak river flow observed at the station had a delay varying from 13 to 41 days among the 12 years (Fig. 6). On average, the delay was ~22 days. The hysteresis indicates the average travel time for the snowmelt water over the basin to reach the hydrometric station.

Results from the LOO-CV show that the 12 models trained using the 12 sets of n-1 (11 years) samples all achieved a correlation coefficient of $r>0.8$ with the observed peak river flows at a significance level of $p<0.003$, except for the run with year 2013 left-out which had a $r=0.71$ and a significance level of $p<0.014$. The model forecasts for peak river flows based on the 12 models (Fig. 7) had a $r=0.72$ and $MAE=0.14$ mm day^{-1}, or 9.7% of the observed mean Q_{peak} value. Compared with the model trained using data for all the years, the deterioration of the model performance in forecasting the peak river flows from LOO-CV reflects the limited number of data samples due to the short records of GRACE data. As the GRACE observations continue, and with the follow-up mission of GRACE-FO, it will be necessary to recalibrate the model to refine the parameter values and to increase the model robustness for peak river flow estimates and flood forecasting.

The impact of measurement and leakage error in the GRACE TWS on the model results was investigated by running the model with TWS adjusted by the error either at the pre-winter time (for baseflow) or at the spring breakup (for surface runoff). The impact of the TWS error on the peak river flow estimates was found to be mostly under $MAE=0.06$ mm day^{-1}, or 4% of the mean Q_{peak} value, which is substantially lower than the modelling error. Compared with that for the Red River basin, the impact of the GRACE TWS error on the modelling results is much smaller for the Mackenzie River basin. This is mainly due to the facts that for the Mackenzie River basin (1) the error in GRACE TWS is small (see Section 2) due to its large area and (2) the snow amount is much larger and the peak river flow is less sensitive to S_b than the Red River basin as discussed above.

The model showed a relatively lower correlation coefficient with observed peak flows for the Mackenzie River than that for the Red River (Wang and Russell, 2016). This is not surprising as the Mackenzie River basin is huge and very complex in physiographic and climate conditions. Sub-basins in the Mackenzie River basin have varying flow regimes as detailed in Woo and Thorne (2003). For instance, most of the rivers in the southern basin and at low altitudes peak in early May, but in rivers at higher latitudes and high altitudes, where snowmelt is delayed, spring peaks occur later. In glacierized basins, the ablation of glaciers intensifies in the summer and this, together with snowmelt at high elevations, prolongs the high flows into summer. Large lakes and reservoirs are highly effective in providing large storage capacities to reduce high flows. The Mackenzie River, although it exhibits essentially a subarctic nival regime of snowmelt-induced peak flow, is a combination of many varying flow regimes of its sub-basins. Our model is a highly simplified representation of this complex system. In contrast, the physiographic and climate conditions for the Red River basin are much more monotonous. Secondly, the small interannual variations in snow amounts and peak flows of the Mackenzie River basin increase the impact of input data errors (Q and T_a) on the model results and impose challenges on robust model calibration and validation.

Several water processes during the period from spring breakup to peak river flow, such as rain, evapotranspiration, soil thaw-induced surface infiltration or groundwater recharge, as well as lake storage and river ice dynamics, may significantly affect the magnitudes of peak river flows. The impacts of these processes are not explicitly included in the model and could be important sources in the modelling errors. In particular, substantial rain events over the snowmelt period can significantly affect peak river flows. The soil thaw-induced surface infiltration and groundwater recharge may have relatively small impacts, as the water flow is mainly from south to north and the lower river sub-basin is usually in a frozen state when the melted water from south arrives. Moreover, recharge of groundwater during this time period (if there is any) would lead to an increase in baseflow, which offsets the impact of reduction in surface runoff due to infiltration of the snowmelt water. The calibration of the model using actual peak river flows also reduces the impact of this process on the peak river flow modelling. Nevertheless, explicitly including the above-mentioned water processes during the period from spring breakup to peak river flow in the model needs to be further studied.

One drawback of our method is that it does not recognise the spatial variations of snow amounts due to the large footprint of GRACE data. Consequently, the model cannot address the spatial snowmelt and water flow patterns within the basin which is expected to largely determine the hysteresis between peak snowmelt and peak river flows and to further reduce the modelling errors in peak flow forecasts. Nevertheless, as a case study with the available data and through comparison with the results from other basins, we have demonstrated encouraging results on peak river flow modelling and flood forecasting based only on GRACE TWS and temperature data. In practice, our GRACE-based method can be used in combination with other available data and methods to help improve the accuracy in peak flow or flood forecasts.

5. Summary

The peak river flow for the Mackenzie River is modelled in this study using GRACE satellite observations and temperature data, which advances the applications of space-based time-variable gravity measurements in cold region flood forecasting. The model estimates peak river flow by simulating peak surface runoff from snowmelt and the corresponding baseflow. The model closely estimated the observed values at a downstream hydrometric station. The results also revealed that on average the travel time for the snowmelt water to reach the hydrometric station is about 22 days. The major driver for determining the peak flow was found to be atmospheric temperature variations. Compared with the Red River basin, the results highlight that the Mackenzie River basin has relatively high water storage and water discharge capability, and low snowmelt efficiency per unit temperature. Our GRACE-based approach for basin-scale snowfall estimation is independent of in situ measurements and largely eliminates the limitations and uncertainties present with traditional approaches. The results show that the GLDAS snowfall amount in our study region is about 20% lower than our GRACE-based estimate. The model is relatively simple and only needs data inputs of GRACE and atmospheric temperature observations for peak flow or flood forecasting. The model can be readily applied to other cold region basins, and could be particularly useful for regions with minimal data. In practice, this GRACE-based method can be used in combination with other available data and methods such as real-time flow data and flow routing models to help improve the accuracy in river flood forecasting, and develop reservoir operation procedures for flood and water resources management.

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Table 1. Baseflow model calibration and performance

a ($\times 10^{-3}$ day^{-1})	b (mm)	Correlation (r)	Significance (p<)	Average winter baseflow (mm)	Mean Absolute Error (MAE)		Nash–Sutcliffe model efficiency coefficient (E)
					(mm)	(%)	
1.07	-195.9	0.73	0.007	48.77	2.37	4.9%	0.53

Table 2. Snowmelt model calibration and model performance for peak surface runoff estimation

α (mm °C^{-1})	β (°C)	Correlation (r)	Significance (p<)	Average peak surface runoff (mm day^{-1})	Mean Absolute Error (MAE)		Nash–Sutcliffe model efficiency coefficient (E)
					(mm day^{-1})	(%)	
17.0	2.1	0.82	0.001	1.26	0.10	7.6	0.50

Table 3. Model performance for peak river flow estimation

Correlation (r)	Significance (p<)	Average peak river flow (mm day^{-1})	Mean Absolute Error (MAE)		Nash–Sutcliffe model efficiency coefficient (E)
			(mm day^{-1})	(%)	
0.83	0.001	1.46	0.1	6.5	0.51

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Figure Captions:

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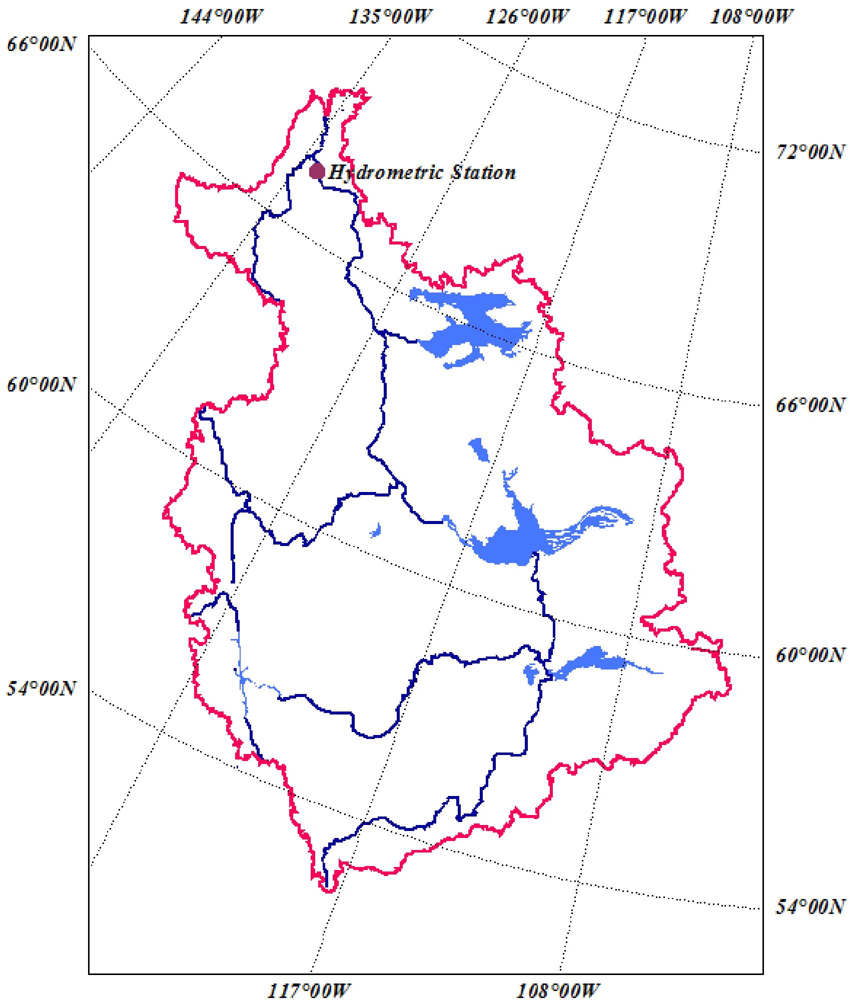

Figure 1. Map for the Mackenzie River basin and the hydrometric station used in this study.

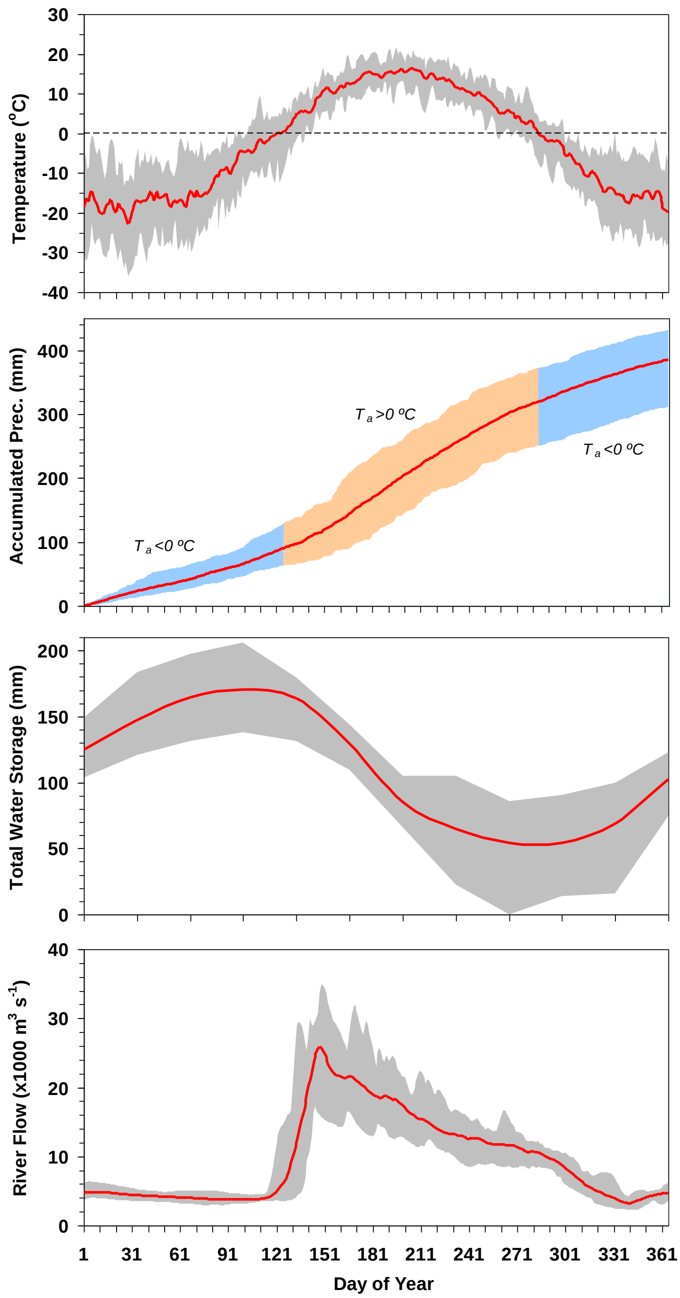

Figure 2. Hydroclimate characterization of the Mackenzie River basin (from top to bottom: air temperature, accumulated precipitation, total water storage change, and river flow. Lines represent the averages over the study period of 2002-2014. Shadowed areas represent the maximum/minimum variation ranges).

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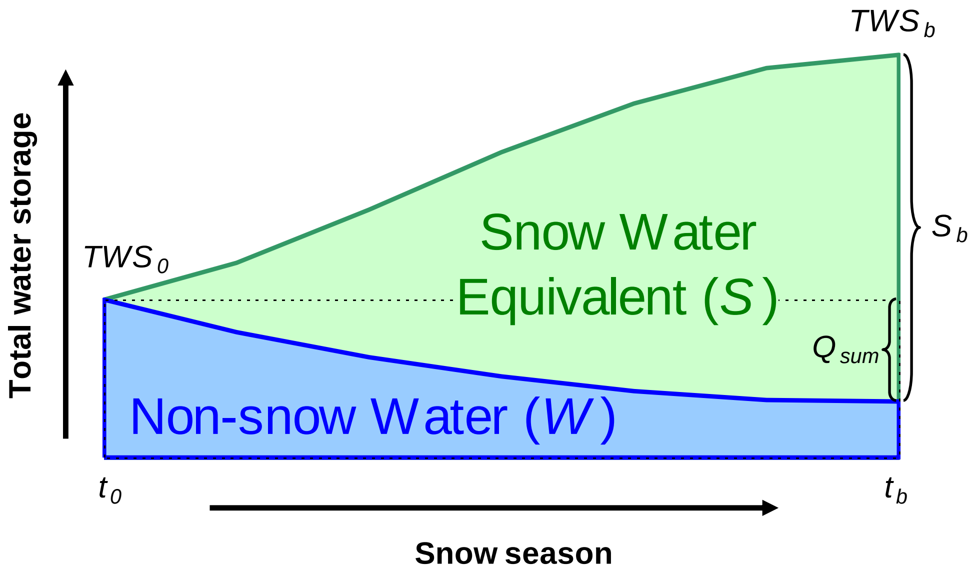

Figure 3. Diagram showing snow accumulation and non-snow water storage change in winter. (t_0 and t_b: start and breakup of snow season; TWS_0 and TWS_b: total water storage at t_0 and t_b; W: non-snow water; S_b: Snow Water Equivalent at t_b; Q_{sum}: total river discharge in snow season).

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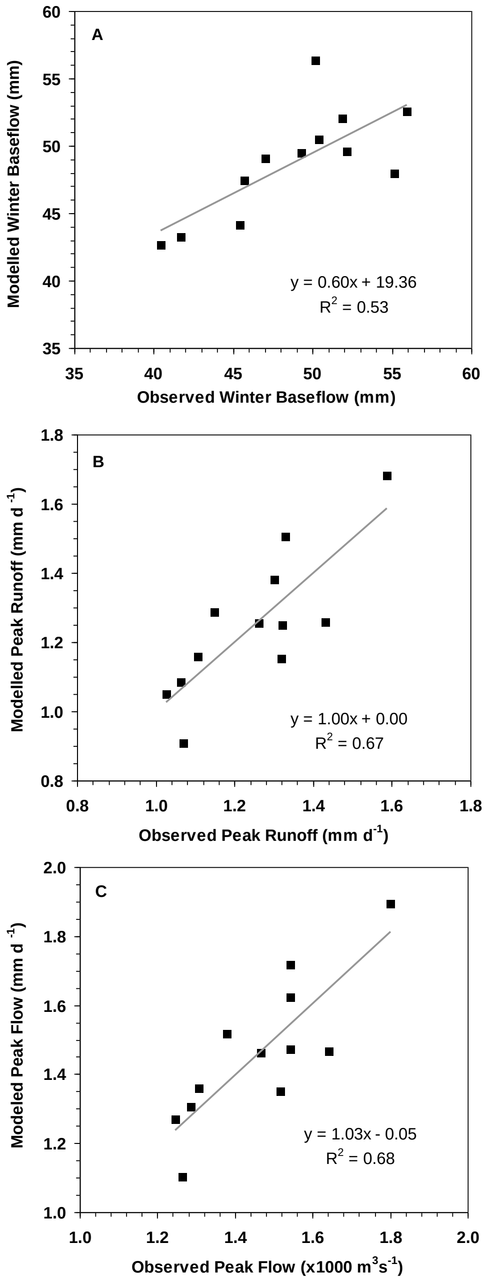

Figure 4. Comparisons of modelled vs. observed total baseflow in winter (A), peak surface runoff (B) and peak river flow (C).

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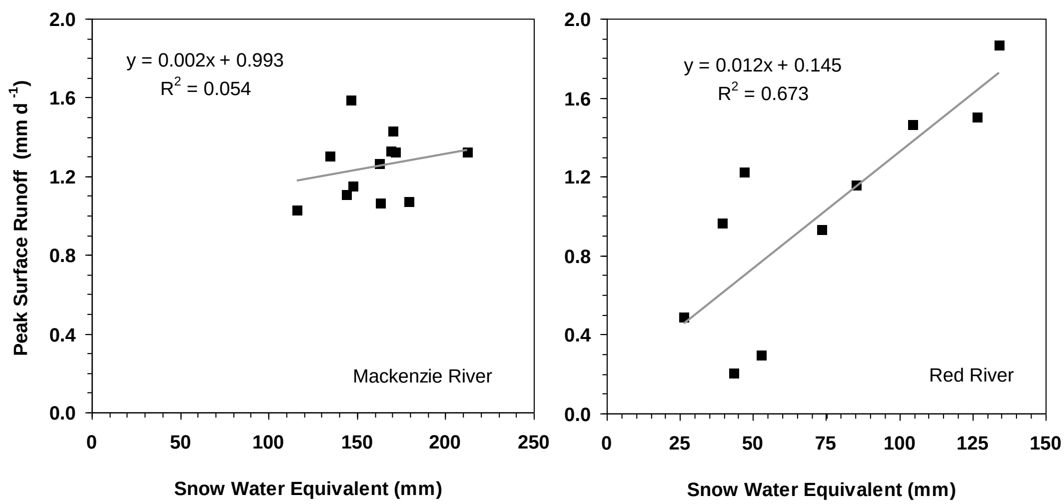

Figure 5. Peak surface runoff vs. snow water equivalent at spring break-up for the Mackenzie River basin (left) and the Red River basin (right).

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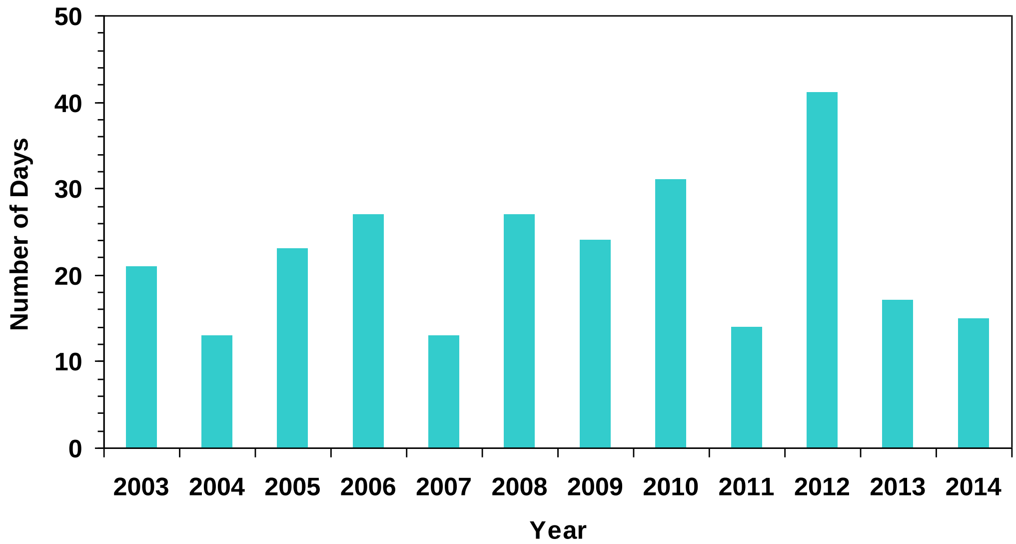

Figure 6. The travel time for snowmelt water from the basin to reach the hydrometric station.

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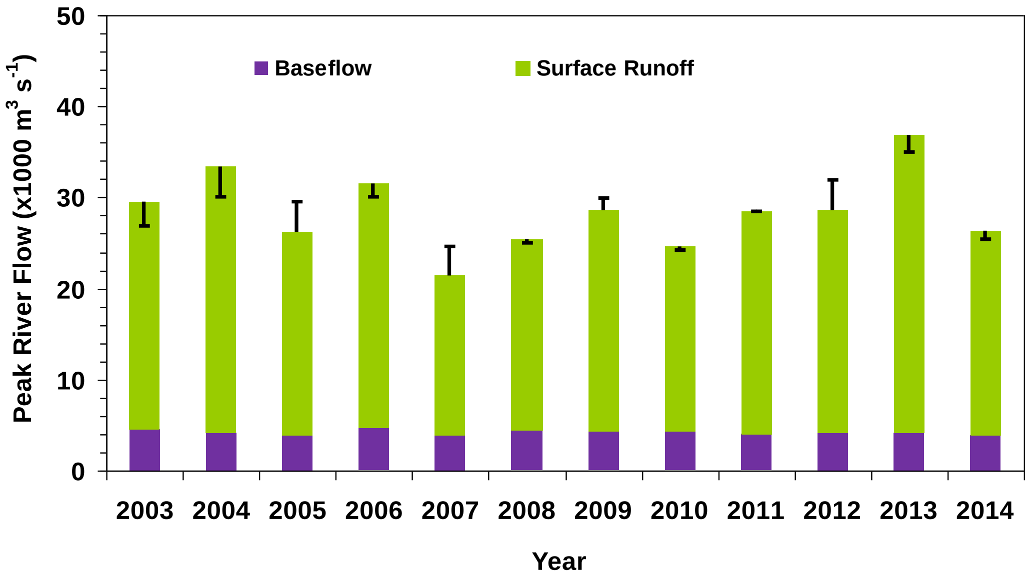

Figure 7. The leave-one-out cross-validation (LOO-CV) modelling results for peak river flow forecast. The error bar is the difference between forecasted and observed peak river flows.