Streamflow hydrology in the boreal region under the influences of climate and human interference

Ming-ko Woo^{1,*}, Robin Thorne¹, Kit Szeto² and Daqing Yang³

¹*School of Geography and Earth Sciences, McMaster University, Hamilton, Ontario, Canada L8S 4K1*

²*Climate Research Division, Environment Canada, Toronto, Ontario, Canada M3H 5T4*

³*Water and Environmental Research Center, University of Alaska, Fairbanks, AK 99775, USA*

The boreal region has a subarctic climate that is subject to considerable inter-annual variability and is prone to impacts of future warming. Climate influences the seasonal streamflow regime which typically exhibits winter low flow, terminated by spring freshet, followed by summer flow recession. The effects of climatic variation on streamflow cannot be isolated with confidence but the impact of human regulation of rivers can greatly alter the natural flow rhythm, changing the timing of flow to suit human demands. The effect of scenario climate change on streamflow is explored through hydrological simulation. Example of a Canadian basin under warming scenario suggests that winter flow will increase, spring freshet dates will advance but peak flow will decline, as will summer flow due to enhanced evaporation. While this simulation was site specific, the results are qualitatively applicable to other boreal areas. Future studies should consider the role of human activities as their impacts on streamflow will be more profound than those due to climate change.

Keywords: boreal region; streamflow regime; human modification of flow; climate change; climate variability

1. INTRODUCTION

The bulk of freshwater that enters the polar seas is supplied directly or indirectly by rivers that drain the boreal region. The largest contribution comes from the Ob ($394 \text{ km}^3 \text{ yr}^{-1}$), the Yenesei ($580 \text{ km}^3 \text{ yr}^{-1}$), the Lena ($528 \text{ km}^3 \text{ yr}^{-1}$) and the Mackenzie ($284 \text{ km}^3 \text{ yr}^{-1}$) Rivers, all of which flow directly into the Arctic Ocean, and the Nelson River ($68 \text{ km}^3 \text{ yr}^{-1}$) that discharges to Hudson Bay, thence to the Arctic. The Yukon River ($105 \text{ km}^3 \text{ yr}^{-1}$) flows to the Bering Sea but much of the water is subsequently conveyed by the ocean current to the Arctic Ocean. Freshwater forms a surface layer on the denser saline seawater. Its presence allows ready formation of sea ice, a process that involves latent heat which is dissipated into the atmosphere to affect climatic feedback. The extent and duration of a sea ice cover affects oceanic evaporation, hence the moisture and heat fluxes into the Arctic atmosphere. A change in freshwater input to the Arctic Ocean therefore has global climatic implications beyond the drainage basins from which the water is derived.

The boreal region is sensitive to variations in the climate. It is also an area where many rivers have been regulated (Ye *et al.* 2003), notably for hydroelectric power generation. Both natural and human factors cause variations and changes in the timing and magnitude, hence the seasonal rhythm of river discharge. The region is also considered by most global climate models (GCMs) to be highly prone to

human-induced warming. This can have significant attendant impacts on the environment (ACIA 2005).

Quantitative assessments of streamflow response to climatic variations and climatic change are needed to provide guidance for the formulation of environmental adaptation strategies and to facilitate rational development of the North. It is therefore the purpose of this paper to examine the seasonal rhythm of streamflow as related to the climate and the effects of river regulation and to assess the impacts of climatic variability and change on river discharge.

2. DATA AND METHODS

The boreal zone is sparse in climatic and hydrometric data (Lammers *et al.* 2001) though there are sufficient streamflow-gauging stations to provide samples across the circumpolar ring of boreal forests. Reanalysis data from major climate centres offer a broad picture of the boreal climate. Together with the application of macroscale hydrological models to simulate streamflow response to scenarios of future climates, we can piece together the mosaic of flow under the present and projected climatic conditions.

Several basins with areas of 10^3 – 10^5 km^2 provide examples for this study (figure 1). Streamflow data are obtained from HYDAT for Canada, <http://nwis.water-data.usgs.gov> for Alaska, and <http://www.r-arcticnet.sr.unh.edu/v3.0/main.html> for Russia and Scandinavia. Climate information is taken from NCEP reanalysis (Kalnay *et al.* 1996). Streamflow simulation uses ERA40 data, a global reanalysis product from the European Centre for Medium-Range Weather Forecasts (see website http://data.ecmwf.int/data/d/era40_daily/).

* Author for correspondence (woo@mcmaster.ca).

One contribution of 12 to a Theme Issue 'The boreal forest and global change'.

Figure 1. Circumpolar boreal region and location of rivers included in this study.

Scenarios from the GCM of the Canadian Centre for Climate modelling and analysis (CCCma) are used for flow simulation under climate change. Scenario data are obtained from website http://cera-www.dkrz.de/IPCC_DDC/SRES/index.html.

Streamflow simulation was performed using the SLURP (Semi-distributed Land Use-based Runoff Processes) model, v. 12.2, which has been used successfully in a number of macroscale hydrological studies in cold regions (Kite *et al.* 1994; Barr *et al.* 1997). The model requires a basin to be subdivided into a number of aggregated simulation areas or ASAs, each encompassing a number of land cover types characterized by a set of parameters. Hydrological simulation comprises: (i) a vertical component consisting of daily surface water balance and flow generation from several storages and (ii) a horizontal component of flow delivery within each ASA and channel routing to the basin outlet. Details on simulation using SLURP, applied to a subarctic basin in Canada (Liard River), can be found in Thorne & Woo (2006).

3. CLIMATE OF THE BOREAL REGION

Climate exerts a commanding influence on hydrology, including the regime of streamflow. The boreal climate is a result of complex interplay among such factors as the large seasonal contrasts in solar input, a wide spectrum of disturbances in the mid- and high-latitude air streams and physiography within or adjacent to the boreal zone. Broadly speaking, the region has a continental subarctic climate. During the long winter, there is little solar energy input, and radiation deficit at the surface often results in extremely low temperatures. Low moisture content in the cold air yields light precipitation. However, sublimation loss is limited under stable atmospheric conditions and the snow can accumulate with little interruption for five or more months until spring arrives.

Long hours of solar heating as solstice approaches and abundant surface moisture from spring snowmelt create conditions favourable to evapotranspiration and the development of convective storms in the summer. Moisture recycling is important in producing warm-season precipitation. For example, close to or more

Figure 2. Linear correlation maps of winter (DJF) temperatures (T) and precipitation (P) from the NCEP reanalysis with corresponding Pacific–North America (PNA) and Arctic Oscillation (AO) teleconnection indexes for the period 1950–2000: (a) PNA and T, (b) AO and T, (c) PNA and P and (d) AO and P.

than half of the summer precipitation is derived from local evaporation over western boreal Canada and Siberia, respectively. Furthermore, the northern tree-line in North America often demarcates the Arctic frontal zone (Serreze *et al.* 2001), and the polar front typically migrates northward to the southern fringe of the boreal zone during the warm season. Synoptic activities and convective storms bring much summer precipitation to the areas.

Within the boreal region, there is considerable variation in the climate due to topographic influence, proximity to large water bodies and position of large-scale atmospheric features such as frontal zones and storm tracks. For example, Atlantic storms often bring abundant winter precipitation to the boreal forests of eastern Canada and northwestern Europe, but distance from the ocean or blocking effect of the Cordillera deprive Siberia or western Canada of heavy winter snowfall. Although the Cordillera limits atmospheric moisture transfer from the Pacific to the boreal zone in western Canada, the release of latent heat associated with orographic precipitation on the windward slopes enhances sensible heat transport to the leeward locations (Szeto *in press*). This effect renders the winters in this area warmer than, say, most of Siberia.

The boreal climate also exhibits large temporal variability forced by low-frequency variations in the large-scale atmospheric circulation, especially during the cold season. Different areas are affected by different circulation regimes (figure 2): northwestern North America is linked to regimes such as the Pacific–North America pattern on the seasonal and inter-annual time scales or the Pacific Decadal Oscillation (Mantua *et al.* 1997) on the decadal scale; and eastern Canada and

northwestern Europe are affected by the Arctic Oscillation (Wallace 2000) or its Atlantic manifestation, the North Atlantic Oscillation. The physical environment of the region can significantly modify the climatic responses to these atmospheric forcings. Interaction of the north Pacific airflow with the Cordillera, for instance, amplifies the winter temperature response in the western Canadian boreal zone to render it the most variable in the world (Szeto *in press*). Enhanced response to changes in the large-scale conditions has yielded some of the strongest climate change signals (Serreze *et al.* 2000; Lucarini & Russell 2002), especially the observed winter warming of northwestern North America and west and central Siberia and cooling over eastern Canada (figure 3).

The precipitation signal, however, is less pronounced though there are indications of slight reduction in winter precipitation over the last 50 years (figure 3). With notable observed climatic variability and change, the boreal region may be an appropriate test bed for examining streamflow response to the climate.

4. STREAMFLOW REGIMES

Streamflow regime is the average seasonal rhythm of river discharge. It reflects the hydrological processes responsible for the production, loss and storage of water in the river basin in which streamflow is generated. Rivers in the boreal region exhibit regimes in response to natural processes as well as to modifications by human activities. We selected examples from rivers in the circumpolar boreal zone to illustrate. Figure 4 plots the monthly discharge of these rivers for a 37–44-year period. For a particular month, each river can produce large inter-annual differences, particularly for the seasons with high flows. Such flow variability may be indicative of fluctuations in the climate and differential release of run-off held in storage as water, snow or ice.

5. NATURAL FLOWS

Rivers in the boreal region typically show a subarctic nival regime as described by Church (1974) in which snowmelt is the principal process that yields the bulk of the total run-off. The Missinaibi River in northern Ontario, Canada, is representative of this seasonal flow pattern (figure 4). In view of the winter climate with prolonged sub-freezing temperatures, snow accumulation is seldom interrupted by melt events. The rivers acquire an ice cover (Prowse & Ferrick 2002) and below the ice, low flow is maintained by groundwater discharge. Snowmelt and river ice break-up in the spring is quickly followed by a rise in streamflow. Like many rivers in the boreal region, the Missinaibi flows from south to north so that both snowmelt contribution and river ice break-up progress northward with the season. Thus, the freshet often begins in April and reaches a peak in May. Afterwards, flow declines as rainfall input is exceeded by evaporation loss and basin storage recharge, until the autumn when frontal storms deposit rainfall in excess of the declining evaporation to generate the secondary high flows. The commencement of winter in December returns to a low-flow season. Latitude affects the timing

Figure 3. (a) Linear trends (K yr^{-1}) of winter (DJF) temperatures from NCEP analysis between 1960 and 2000. (b) Correlations of NCEP DJF precipitation time series with time (in years) to give a qualitative indication of winter precipitation trends (i.e. positive values indicate positive trends and vice versa); correlation is provided instead of the trend values because the amount of precipitation in the boreal zone typically has a very large range, with overwhelmingly high values along storm tracks and in mountainous areas.

Figure 4. Monthly flows of unregulated boreal rivers showing several types of streamflow regimes: (i) subarctic nival regime—Missinaibi, Upper Liard and Tana, (ii) proglacial regime—Tanana, (iii) prolacustrine regime—Lockhart, Great Bear, (iv) nival-pluvial regime—Rupert. Dashed lines represent the monthly mean for the first 20 years. Solid lines are for the last 20 years.

of hydrological events. The Rupert River in Quebec, Canada, being further north, has a later high flow than the Missinaibi due to delayed snowmelt (figure 4). Furthermore, the basin receives much rain in the summer that amply compensates evaporative loss so that the summer flow decline is less intense than in the Missinaibi. This adds a pluvial element to the predominantly nival regime. Not all rivers have a secondary high flow in the autumn as its occurrence relies on frequent passages of autumnal storms. For example, the Tana River that flows between Finland and Sweden does not possess this feature (figure 4) since most of the rainfall falls in the summer months (Dankers & Christensen 2005).

The Upper Liard River in northwestern Canada has high flows delayed for a month (in June) when compared with the Missinaibi because it is further north. Its mountainous setting also gives rise to altitudinal effects in that snowmelt progresses from the valleys to the higher elevations so that melt run-off is often spread over a two-month period (Woo & Thorne 2006).

The presence of lakes and glaciers gives rise to variations to the standard subarctic nival regime (Woo 2000). The Tanana River in Alaska is fed by glaciers of the Alaska Range. It has the typical winter low flow under an ice cover and a spring freshet caused by snowmelt in May and June (figure 4). Instead of a declining summer flow, however, discharge continues to increase in the summer as run-off from glacier melt feeds the headwater streams. These features produce a regime with a proglacial flavour. Lakes abound in many boreal plains and Precambrian bedrock terrain. Their ability to store and then release water buffers the streamflow from extremes. Lockhart River in Northwest Territories, Canada, flows through a chain of lakes amid the bedrock upland of the Canadian Shield. Although its latitude is comparable with the Upper Liard River, the Lockhart has more subdued and later high flows (figure 4). While the lakes and the river system also acquire an ice cover in the winter, there is a steady release of flow from the lakes to sustain higher low flows than other boreal rivers without the lake influence. An extreme case of streamflow with a lacustrine regime is the Great Bear River issued from Great Bear Lake, Northwest Territories (figure 4). This lake, being one of the 10 largest in the world, dominates the drainage basin and its storage alters the streamflow regime to the extent that annual high flows are delayed until August and September, at least three months after the regional snowmelt.

There is a growing literature on the streamflow trends of rivers that drain to the Arctic Ocean. Peterson *et al.* (2002) reported an increase in the flow of the six largest Eurasian rivers in the past decades and related this phenomenon to changes in both the North Atlantic Oscillation and the global warming. Yang *et al.* (2003, 2004) and Ye *et al.* (2003) reported annual streamflow increases of 3–7% over the past six decades in the large Siberian rivers. Lammers *et al.* (2001) and Serreze *et al.* (2003) also found that most large Siberian rivers show an increasing trend in run-off, especially during winter and spring. The causes for these changes are unclear but the spring discharge increase has been attributed to an earlier snowmelt associated with warmer spring conditions, and changes in winter streamflow are

perhaps associated with an increase in the depth of the active layer (seasonally frozen and thaw zone above the permafrost) under a warming climate (Yang *et al.* 2002). McClelland *et al.* (2004) suggested that increased precipitation was a key factor in streamflow augmentation, but not permafrost thaw or forest fire. Déry & Wood (2005) noted a flow reduction from northern continental Canada, including the Hudson Bay drainage. Abdul Aziz & Burn (2006) examined the monthly flows of the Mackenzie basin and found a diversity of trends for its sub-basins, notably an increase in the winter flows and weakly decreasing flows in summer and late autumn. Woo *et al.* (2006) commented that streamflow response to climatic forcing is complicated by location, topography and storage, and what is considered as a climatic trend may instead be a shift in the climatic regime. Placed in the context of these discussions and in view of the limited record lengths, the flow data of the selected boreal rivers in North America were partitioned into two periods of approximately equal length (though not of the same years) for comparison of their means. No test of statistical significance was performed and only the signs were noted to offer a general sense of streamflow change. Eurasian rivers were excluded as most of them are too heavily regulated to manifest the effects due to natural processes (see §6 below).

Both the Tanana and the Upper Liard River in northwestern North America showed an increase in winter low flow but a slight reduction in snowmelt run-off. The Upper Liard continued to have a flow decrease in the summer but the Tanana experienced an increase in summer flow. The difference may be attributed to the occurrence of more frequent warm summers since 1980, which led to greater evaporation loss, but for the Tanana this was compensated by larger glacier melt. One may suggest the influence of a shift in climatic regime, as El Niño-Southern Oscillation (ENSO) events became more frequent after the mid-1970s. The Great Bear River that lies to the east of the Cordillera showed a general reduction in flow, especially between June and September. However, the Lockhart River, approximately 1200 km east of the Upper Liard and 700 km southeast of the Great Bear, had an increase in mean monthly flows, notably during the high-flow months of July to October. Such mixed signals cannot be related easily to the variation of the regional climate. Further east in Ontario and Quebec, the Missinaibi and the Rupert Rivers showed the opposite tendency of streamflow decline in most months, possibly confirming the decreasing flow trend indicated by Déry & Wood (2005). Flow reduction in May for the Missinaibi should be considered in conjunction with its flow increase in April. This is probably related to an earlier arrival of snowmelt that augments the April flow but diminishes the magnitude of the freshet, a feature that has been noted by Burn & Hag Elnur (2002) for some rivers in Canada.

6. REGULATED REGIMES

Streamflow regulation often alters the natural seasonal discharge cycle (Yang *et al.* 2004). We selected three river systems, two in Siberia and one in Canada, to

Figure 5. Monthly flow at three stations (a) Chernyshevskyi Station (136 000 km²; grey shade, 1959–1967, black shade, 1968–1990), (b) Suntan (202 000 km²; grey shade, 1936–1967, black shade, 1968–1999) and (c) Hatyrik-Homo (452 000 km²; grey shade, 1936–1967, black shade, 1968–1999) along Vilui River before and after damming at Chernyshevskyi Station. There is evidence of diminishing regulation effect downstream of the dam.

illustrate the flow regulation effects. The example of Vilui River in the Lena basin shows a partial recovery to the natural flow regime downstream of a dam. This river originally manifested a subarctic nival regime. After a dam was built in 1967, winter flow at the Chernyshevskyi Station increased by approximately 400–600 m³ s^{−1} (approx. 10–100 times of the pre-dam discharge) during November–April. Streamflow was reduced by 1400 m³ s^{−1} (80%) in May and 2300 m³ s^{−1} (70%) in June. July discharge increased by 300 m³ s^{−1} (40%), but small changes (less than 25%) were observed during August to October (figure 5a). Approximately 350 km downstream of the Chernyshevskyi dam at Suntan Station, similar changes in mean monthly flow were maintained (figure 5b). However, at the Hatyrik-Homo Stations located 900 km downstream of the dam, the difference in mean May discharges was substantially reduced due to increased run-off contribution from the unregulated areas within the Vilui valley. The impact of reservoir regulation is most obvious during winter months and also in June when its flow was reduced by 2200 m³ s^{−1} or 28% (figure 5c).

The effects of flow regulation may sometimes propagate downstream for a considerable distance.

Figure 6. Changes in streamflow regime before and after reservoirs were in operation at: (a) Kolyma River at Ust'-Srednekan station in Siberia (99 400 km²) (grey shade, 1933–1986, black shade, 1986–2000) and (b) Peace River at Hudson Hope in western Canada (69 900 km²) (grey shade, 1950–1966, black shade, 1968–1984).

An example is the Ust'-Srednekan station in Siberia, located approximately 1500 km downstream of a dam built along the main valley of the Kolyma River basin (Petrov & Losev 1976). This rock-filled dam measures 130 m high and 780 m long and impounds a reservoir with a surface area of 441 km². After the reservoir was filled between 1986 and 1990, the station experienced a large winter (December–April) flow increase of approximately 200 m³ s^{−1} and a reduction in June peak flow by 1330 m³ s^{−1}. As a result, the monthly hydrograph changed significantly: summer flows became lower but winter flows increased, thus moderating the seasonal differences (figure 6a).

In an extreme case, as exemplified by the Peace River which is a tributary of the Mackenzie River, there is a reversal of the high- and low-flow periods due to flow releases regulated to meet the seasonal demands for hydroelectric power. Before regulation, the river exhibited a nival flow regime (figure 6b). After the Bennett Dam came into operation in 1968, winter emerged as the high-flow period whereas the spring was a low-flow season (Peters & Prowse 2001). Although the flow has become relatively more uniform through the year, human interference with the outflow can occasionally produce extreme events. Woo & Thorne (2003) noted that an artificial release of water from the reservoir in 1996, in concert with high-precipitation downstream in 1997 that was withheld in storage by Great Slave Lake further downstream and then discharged from the Lake in 1998, produced an unprecedented peak flow event for the Mackenzie River.

7. EXPECTED CLIMATE CHANGE

Human-induced global warming is expected to cause long-term changes in the climate. In the coming

Figure 7. Liard basin showing air temperature of (i) winter (November–March), (ii) spring (April–June), (iii) summer (July–August) and (iv) autumn (September–October) as depicted by the ERA40 for: (a) the present (1961–1990) climate, (b) temperature change for 2050 and (c) for 2100, according to CCCma for the B2 scenario.

century, climate change may lead to a rise in the mean annual temperature of the boreal zone by 3–4°C, though there are discrepancies among the predictions from various climate models or from different greenhouse gas emission scenarios assumed for the future. The increases are stronger during the winter and spring than in other seasons. Annual precipitation in the broad boreal zone is also predicted to increase by approximately one inter-annual standard deviation above their current values, with the strongest projected increase occurring in winter and spring.

The IPCC Special Report on Emissions Scenarios has suggested four qualitative greenhouse gas emission drivers under future developments that may be more economically oriented or environmentally cognizant. For our study, we chose the more conservative B2 scenario in which local solution to sustainability is applied to a heterogeneous world. Under such a scenario, various GCMs have produced their modelled climate change conditions, from which we selected the results yielded by the CCCma for our study of streamflow response.

Rather than investigating streamflow changes across the entire boreal zone, we chose the Liard basin (area 275 000 km²) in northwestern Canada for an in-depth study. This basin straddles high mountain chains of the western Cordillera and the plateaus and lowlands of the Interior Plains in the east, with an elevation range from 140 to 2700 m, offering a variety of topography and local climates. Figures 7 and 8 show the spatial distributions of air temperature and precipitation for the four seasons of winter (November–March), spring (April–June), summer (July–August) and autumn

(September–October) as depicted by the ERA40 for the present (1961–1990) climate. Also shown are the changes in temperature (°C) and precipitation (mm) according to the B2 scenarios provided by CCCma for 2050 (average of 2021–2050) and for 2100 (average of 2071–2100).

Under the present climate, low winter temperatures of less than –10°C prevail throughout the basin. Winter is the wettest season, particularly in the western part where orographic precipitation deposits much snow. Spring arrives with above-freezing temperatures that average less than 5°C in the western half of the basin and 5–10°C in the east. Orographic precipitation remains significant at high elevations but there is a basin-wide decrease as the season progresses. Summer can warm above 15°C in the low-lying areas, but is cooler in the mountains. Precipitation increases in the autumn as frontal storms become more frequent. The southwestern highlands receive the most precipitation, particularly from the Pacific airflow that tracks across the mountains.

With climate change comes the expectation of warming of the basin, with greater warming for 2100 than 2050, especially for winter and spring, and for the western mountain areas. CCCma data indicate a large reduction of winter precipitation (down by –40 mm) in the southeastern corner, but an increase (up by +30 mm) in the spring for the basin, particularly its southern sector. Summer and autumn precipitation will be little changed, being approximately ± 10 mm of the present, though there is a switch in the autumn, from a precipitation decrease in 2050 to a slight increase in 2100.

Figure 8. Liard basin showing: (a) precipitation of the four seasons ((i) November–March, (ii) April–June, (iii) July–August and (iv) September–October) as depicted by the ERA40 for the present climate, (b) precipitation change for 2050 and (c) for 2100, according to CCCma for the B2 scenario.

8. STREAMFLOW RESPONSE TO CLIMATE CHANGE: AN EXAMPLE

Hydrologic simulation was performed by the SLURP hydrological model run on daily time steps. The results were averaged or summed at monthly intervals for presentation purposes. The simulation assumed no change in land cover and soil conditions under a natural (little human disturbance) setting. Gridded (at a resolution of 2.5° latitude–longitude) daily temperature and precipitation from ERA40 for 1961–1990, covering the entire Liard basin, were used as inputs to simulate streamflow for the present climate. Temperature and precipitation changes as depicted by CCCma were superimposed onto the 1961–1990 daily data to generate synthetic daily temperature and precipitation series for 2050 and 2100, for use as inputs to SLURP to simulate future streamflow for Liard River at its confluence with the Mackenzie River at Fort Simpson.

Using 30 years of simulated values for each set of conditions (present, 2050 and 2100), it is possible to estimate the exceedance probability (probability that a particular magnitude would be exceeded) of various flow attributes. A plot of the exceedance probability of the annual flow (figure 9b) indicates that under the future climates, the Liard discharge will not be altered. This is due to the incremental precipitation being able to compensate for the enhanced evaporation under climate warming. In addition, climate change will affect all major features of the subarctic nival regime including high and low flows, as the timing and the basin water balance will be modified.

One notable feature is the increase of winter low flow, especially at the end of the century. This is due to sporadic occurrences of rising flows generated by winter rain or occasional winter melt events that become possible under a warming climate. Such events have been observed in the temperate latitudes of eastern Canada and northeastern United States (Beltaos 2002) and are expected to become a regular feature in the boreal region. Warmer spring seasons lead to earlier arrival of snowmelt run-off so that the starting date of spring freshet is advanced (figure 9). This leads to increased flow in April, accompanied by a reduction in peak flow in May. The change is more marked for 2100 than for 2050 as warming intensifies. Such shifts in the probability distributions are attributed to a combination of reduced winter snow accumulation where winter rainfall events have increased (by 3% for 2050 and 13% for 2100) and intermittent winter melt, and an earlier melt season that spreads the basin snowmelt over a longer period. The latter phenomenon can be seen even under the present climate in which a late melt year withholds the bulk of snow until a sharp increase in melt energy synchronously releases the snow at all elevations; but an early melt year depletes the snow gradually so that run-off becomes less concentrated (Woo & Thorne 2006). Zhang *et al.* (2001) similarly found an earlier warming in spring leading to earlier and more gradual snowmelt in a number of Canadian rivers.

An early arrival and termination of the snowmelt period is complemented by an extension of the summer

Figure 9. Exceedance probability of several streamflow characteristics of the Liard River for the present climate and under the B2 scenario for 2050 and for 2100: (a) date of arrival of spring freshet (hydrograph rise), (b) annual flow, and monthly flows of (c) April, (d) May and (e) June (black filled circle, present; grey filled circles, 2050; open circles, 2100).

season, with an attendant increase in evaporation that is enhanced by the temperature increase postulated for the summer. Although annually, increased evaporation loss is compensated by total precipitation, higher evaporation in the summer is not offset by summer rain which is not expected to undergo significant changes. Summer streamflow will consequently decrease. Autumn months have the least change in flow and the secondary annual high flow is still maintained under the future climates. Overall, the subarctic nival regime will be preserved in the Liard River, but there will be a rearrangement in the timing and the magnitude of its seasonal discharge.

While the result of this simulation is location specific, the approach is applicable to other large catchments. For the modelling of large basin areas, the topography, vegetation and soils are necessarily generalized. Yet, future studies may include in their simulations the change in land cover and soil

properties. We note further that, for this simulation, the basin is assumed to retain its pristine conditions, a situation that is unlikely to be sustained as development pressure will no doubt be exerted on most boreal zones.

9. DISCUSSION AND CONCLUSIONS

The circumpolar boreal region is sensitive to inter-annual variations in its climate, especially in the cold season, and climate model results suggest greater climatic changes for the boreal regions than for most other land parts of the world. Streamflow in the region manifests strong seasonality as a response to the subarctic climate. The prevalent flow pattern is the nival regime in which low flow of the cold winter switches quickly to the snowmelt freshet, followed by summer flow recession as evaporation intensifies. Local conditions give rise to several variants, including the proglacial regime with glacier meltwater augmentation, lacustrine regime in which lake storage moderates streamflow and pluvial modification in areas with high summer rain.

Climate variability signals, clearly manifested in the temperature and less so in the precipitation of the boreal region, cannot be confirmed with certainty in the flow response, even though recent trends can be detected by statistical analyses of individual stations and for particular periods of the year. A host of factors, such as location, topography and basin storage can modify the climatic influence. In contrast, human influence through the impoundment, release and diversion of flow can be readily discerned. These impacts on river discharge usually overwhelm the effects of the climate and must be recognized as an attribute of most boreal river systems.

Hydrological modelling can be used to simulate the effects of climate change on streamflow. The case study of Liard River suggests that, in the absence of major environmental changes in the basin, warmer winters will increase the frequency of rain and intermittent snowmelt, while spring warming can advance the timing of break-up to extend the snowmelt season, leading to less intense freshets. Warmer and longer summers will increase evaporation loss to the detriment of summer flow. These generalities may apply qualitatively to other boreal rivers but hydrological simulations have to be performed for specific areas as the timing and magnitude of flow changes will vary among basins. Future studies should incorporate external variables that include vegetation and ecosystem changes, permafrost and soil dynamics feedbacks, and, most importantly, the roles of human interference as future development will surely stress the boreal zone.

It must be emphasized that modelling is by no means a substitute for observations. Over the past decades, the data collection network in the boreal region has undergone serious attrition worldwide. For an area sensitive to climatic forcing and human development, the monitoring of its climate and streamflow must be maintained, and even intensified.

This study is a contribution of the Mackenzie Global Energy and Water Cycle Experiment (GEWEX) Study supported by the Natural Sciences and Engineering Research Council of Canada. We thank Dr Doug Kane for his helpful comments on the manuscript.

REFERENCES

- Abdul Aziz, O. I. & Burn, D. H. 2006 Trends and variability in the hydrological regime of the Mackenzie River Basin. *J. Hydrol.* **319**, 282–294. (doi:10.1016/j.jhydrol.2005.06.039)
- ACIA 2005 *Arctic climate impact assessment*. Cambridge, UK: Cambridge University Press.
- Barr, A. G., Kite, G. W., Granger, R. & Smith, C. 1997 Evaluating three evapotranspiration methods in the SLURP macroscale hydrological model. *Hydrol. Process.* **11**, 1685–1705. (doi:10.1002/(SICI)1099-1085(19971030)11:13<1685::AID-HYP599>3.0.CO;2-T)
- Beltaos, S. 2002 Effects of climate on mid-winter ice jams. *Hydrol. Process.* **16**, 789–804. (doi:10.1002/hyp.370)
- Burn, D. H. & Hag Elnur, M. A. 2002 Detection of hydrologic trends and variability. *J. Hydrol.* **255**, 107–122. (doi:10.1016/S0022-1694(01)00514-5)
- Church, M. 1974 Hydrology and permafrost with reference to northern North America. In *Proc., workshop seminar on permafrost hydrology* (ed. J. Demers), pp. 7–20. Ottawa, Canada: Secretariat, Canadian National Committee, International Hydrological Decade, Environment Canada.
- Dankers, R. & Christensen, O. B. 2005 Climate change impact on snow coverage, evaporation and river discharge in the sub-arctic Tana Basin, Northern Fennoscandia. *Clim. Change* **69**, 367–392. (doi:10.1007/s10584-005-2533-y)
- Déry, S. J. & Wood, E. F. 2005 Decreasing river discharge in northern Canada. *Geophys. Res. Lett.* **32**, L10 401. (doi:10.1029/2005GL022845)
- Kalnay, E. *et al.* 1996 The NCEP/NCAR 40-year reanalysis project. *Bull. Am. Meteorol. Soc.* **77**, 437–471. (doi:10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2)
- Kite, G. W., Dalton, A. & Dion, A. 1994 Simulation of streamflow in a macroscale watershed using general circulation model data. *Water Resour. Res.* **30**, 1547–1559. (doi:10.1029/94WR00231)
- Lammers, R., Shiklomanov, A., Vörösmarty, C. J., Fekete, B. & Peterson, B. J. 2001 Assessment of contemporary arctic river runoff based on observational discharge records. *J. Geophys. Res.* **106**, 3321–3334. (doi:10.1029/2000JD900444)
- Lucarini, V. & Russell, G. L. 2002 Comparison of mean climate trends in the Northern Hemisphere between National Centers for Environmental Prediction and two atmosphere–ocean model forced runs. *J. Geophys. Res.* **107**, 4269. (doi:10.1029/2001JD001247)
- Mantua, N. J., Hare, S. R., Wallace, J. M. & Francis, R. C. 1997 A Pacific decadal climate oscillation with impacts on salmon production. *Bull. Am. Meteorol. Soc.* **78**, 1069–1079. (doi:10.1175/1520-0477(1997)078<1069:APICOW>2.0.CO;2)
- McClelland, J. W., Holmes, R. M., Peterson, B. J. & Stieglitz, M. 2004 Increasing river discharge in the Eurasian Arctic: consideration of dams, permafrost thaw, and fire as potential agents of change. *J. Geophys. Res.* **109**, D18102. (doi:10.1029/2004JD004583)
- Peters, D. L. & Prowse, T. D. 2001 Regulation effects on the lower Peace River, Canada. *Hydrol. Process.* **15**, 3181–3194. (doi:10.1002/hyp.321)
- Peterson, B. J., Holmes, R. M., McClelland, J. W., Vörösmarty, C. J., Lammers, R. B., Shiklomanov, A. I., Shiklomanov, I. A. & Rahmstorf, S. 2002 Increasing river discharge to the Arctic Ocean. *Science* **298**, 2171–2173. (doi:10.1126/science.1077445)
- Petrov, V. G. & Losev, E. D. 1976 *Dam of the Kolyma hydroelectric power plant. CRREL-TL-563*. Hanover, NH: Corps of Engineers, U.S. Army, Cold Regions Research and Engineering Laboratory.
- Prowse, T. D. & Ferrick, M. G. 2002 Hydrology of ice-covered rivers and lakes: scoping the subject. *Hydrol. Process.* **16**, 759–762. (doi:10.1002/hyp.373)
- Serreze, M. C. *et al.* 2000 Observational evidence of recent change in the northern high-latitude environment. *Clim. Change* **46**, 159–207. (doi:10.1023/A:1005504031923)
- Serreze, M. C., Lynch, A. H. & Clark, M. P. 2001 The summer Arctic frontal zone as seen in NCEP–NCAR reanalysis. *J. Clim.* **14**, 1550–1567. (doi:10.1175/1520-0442(2001)014<1550:TAFZAS>2.0.CO;2)
- Serreze, M. C., Bromwich, D. H., Clark, M. P., Etringer, A. J., Zhang, T. & Lammers, R. 2003 Large-scale hydro-climatology of the terrestrial Arctic drainage system. *J. Geophys. Res. Atmos.* **108**, 8160. (doi:10.1029/2001JD000919)
- Szeto, K. K. In press. On the inter-annual variability of winter temperatures in Northwest Canada. *J. Clim.*
- Thorne, R. & Woo, M.-k. 2006 Efficacy of a hydrologic model in simulating discharge from a large mountainous catchment. *J. Hydrol.* **330**, 301–312. (doi:10.1016/j.jhydrol.2006.03.031)
- Wallace, J. M. 2000 North Atlantic Oscillation/annular mode: two paradigms—one phenomenon. *Q. J. R. Meteorol. Soc.* **126**, 791–805. (doi:10.1256/smsqj.56401)
- Woo, M.-k. 2000 Permafrost and hydrology. In *The Arctic: environment, people, policy* (eds M. Nuttall & T. V. Callaghan), pp. 57–96. Amsterdam, The Netherlands: Harwood Academic Publishers.
- Woo, M.-k. & Thorne, R. 2003 Streamflow in the Mackenzie basin, Canada. *Arctic* **56**, 328–340.
- Woo, M.-k. & Thorne, R. 2006 Snowmelt contribution to discharge from a large mountainous catchment in subarctic Canada. *Hydrol. Process.* **20**, 2129–2139. (doi:10.1002/hyp.6205)
- Woo, M.-k., Thorne, R. & Szeto, K. K. 2006 Reinterpretation of streamflow trends based on shifts in large-scale atmospheric circulation. *Hydrol. Process.* **20**, 3995–4003. (doi:10.1002/hyp.6590)
- Yang, D., Kane, D. L., Hinzman, L. D., Zhang, X., Zhang, T. & Ye, H. 2002 Siberian Lena River hydrologic regime and recent change. *J. Geophys. Res.* **107**, 4694. (doi:10.1029/2002JD002542)
- Yang, D., Robinson, D., Zhao, Y., Estilow, T. & Ye, B. 2003 Streamflow response to seasonal snow cover extent changes in large Siberian watersheds. *J. Geophys. Res.* **108**, 4578. (doi:10.1029/2002JD003149)
- Yang, D., Ye, B. & Shiklomanov, A. 2004 Discharge characteristics and changes over the Ob River watershed in Siberia. *J. Hydrometeorol.* **5**, 595–610. (doi:10.1175/1525-7541(2004)005<0595:DCACOT>2.0.CO;2)
- Ye, B., Yang, D. & Kane, D. L. 2003 Changes in Lena River streamflow hydrology: human impacts versus natural variations. *Water Resour. Res.* **39**, 1200. (doi:10.1029/2003WR001991)
- Zhang, X., Harvey, K. D., Hogg, W. D. & Yuzyk, T. R. 2001 Trends in Canadian streamflow. *Water Resour. Res.* **37**, 987–998. (doi:10.1029/2000WR900357)