Hydrological extremes in a southwestern Ontario river basin under future climate conditions

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Abstract The global climate change may have serious impacts on the frequency, magnitude, location and duration of hydrological extremes. Changed hydrological extremes will have important implications on the design of future hydraulic structures, flood-plain development, and water resource management. This study assesses the potential impact of a changed climate on the timing and magnitude of hydrological extremes in a densely populated and urbanized river basin in southwestern Ontario, Canada. An ensemble of future climate scenarios is developed using a weather generating algorithm, linked with GCM outputs. These climate scenarios are then transformed into basin runoff by a semi-distributed hydrological model of the study area. The results show that future maximum river flows in the study area will be less extreme and more variable in terms of magnitude, and more irregular in terms of seasonal occurrence, than they are at present. Low flows may become less extreme and variable in terms of magnitude, and more irregular in terms of seasonal occurrence. According to the evaluated scenarios, climate change may have favourable impacts on the distribution of hydrological extremes in the study area.

Key words climate change; hydrological modelling; flood; low flow; climate scenario; Canada

Extrêmes hydrologiques dans un basin versant du sud-ouest de l'Ontario sous conditions climatiques futures

Résumé Le changement climatique global peut avoir des impacts significatifs sur la fréquence, l'amplitude, la localisation et la durée des extrêmes hydrologiques. Les extrêmes hydrologiques modifiés auront d'importantes implications sur le dimensionnement des futures structures hydrauliques, le développement des plaines d'inondation et la gestion des ressources en eau. Cette étude estime l'impact potentiel d'un climat modifié sur la répartition temporelle et l'amplitude des extrêmes hydrologiques dans un bassin versant densément peuplé et urbanisé du sud-ouest de l'Ontario, au Canada. Un ensemble de scénarios climatiques futurs est développé grâce à un algorithme de génération de temps, lié aux sorties de modèles climatiques globaux. Ces scénarios climatiques sont ensuite transformés en écoulement de bassin grâce à un modèle hydrologique semi-distribué de la zone d'étude. Les résultats montrent que les écoulements fluviaux maximum à venir dans la zone d'étude seront moins extrêmes et plus variables en termes d'amplitude, et plus irréguliers en termes d'occurrence saisonnière, qu'aujourd'hui. Les étiages peuvent devenir moins extrêmes et variables en termes d'amplitude, et plus irréguliers en termes d'occurrence saisonnière. Selon les scénarios évalués, le changement climatique peut avoir des impacts favorables sur la distribution des extrêmes hydrologiques dans la zone d'étude.

Mots clefs changement climatique; modélisation hydrologique; crue; étiage; scénario climatique; Canada

INTRODUCTION

There is a broad agreement in the international scientific community that global climate change will alter the frequency and magnitude of hydrological extremes. Increased concentration of greenhouse gases in the atmosphere leads to increased air temperatures. Higher air temperatures will in turn intensify the hydrological cycle. The enhanced hydrological cycle will then likely produce extremes different from those

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historically observed. According to the Intergovernmental Panel on Climate Change (IPCC, 2001), the atmospheric concentration of CO_2 has increased from 280 ppm in 1750 to 367 ppm in 1999, and is expected to reach 463–623 ppm by 2050 and 470–1099 ppm by 2100. IPCC further indicates that the global average air temperature has increased over the 20th century by about 0.6 ± 0.2°C, and this increase is the largest of any century during the past 1000 years. Depending on the different emission scenarios, the IPCC projects a further increase of global air temperature in the range of 1–5°C.

Recent scientific literature on the impact of climate variability and change on river flow is voluminous both in the context of observations and projections (see e.g. McKerchar & Henderson, 2003; Paturel *et al.*, 2003; Pongracz *et al.*, 2003; Koutsoyiannis, 2003; Tate *et al.*, 2004; Kundzewicz, 2004; Lindström & Bergström, 2004). The effect of climate change on river flow will largely follow projected changes in precipitation (Arnell, 1999; Pal *et al.*, 2004). Different trends in precipitation are expected in different parts of the world, with a general increase in the Northern Hemisphere and high latitudes (particularly in autumn and winter) and a decrease in the tropics and subtropics in both hemispheres. McGuffie *et al.* (1999) projected increase in the frequency of heavy rainfall events under warmed climate. Changes in precipitation will likely be amplified in basin river runoff. Chiew & McMahon (2002) projected that in wet and temperate Australian catchments the percentage change in runoff can be about twice the percentage change in precipitation, whereas in ephemeral catchments the percentage change in runoff can be more than four times the percentage change in precipitation.

The potential impacts of climate change on hydrological extremes have received considerable attention from hydrologists during the last decade. Many studies suggest that the global warming will increase the frequency and magnitude of extreme hydrological events (Kite, 1993; Boorman & Sefton, 1997; Panagoulia & Dimou, 1997; Gellens & Roulin, 1998; Saelthun *et al.*, 1998; Mirza *et al.*, 1998; Prudhomme *et al.*, 2003; Meehl & Tebaldi, 2004; among others). According to IPCC (2001), flood magnitude and frequency are likely to increase in most regions, and low flows are likely to decrease in many regions of the world. Climate change may also alter the timing of extreme runoff. Cooley (1990) stated that changing the air temperature by only 2–4°C can have a significant impact on snow accumulation and melt rate. Satellite data already show decreases of about 10% in the extent of snow cover since the late 1960s (IPCC, 2001). In fact, in many areas where snowfall is currently an important component of the water balance, snowmelt-induced peak flow is likely to move from spring towards winter (Burn, 1994; Li & Simonovic, 2002; Eckhardt & Ulbrich, 2003).

Because of its location, Canada is projected to experience greater rates of warming than many other regions of the world. According to Lemmen & Warren (2004), changes in Canadian climate will be variable across the country, with the Arctic and the southern and central Prairies expected to warm the most. The average air temperature in Canada has already risen 1.1°C in the past century (Gullet & Skinner, 1992; Koshida & Avis, 1998). Canada has a relative abundance of water, but its resources are not evenly distributed across the country. As a result, most regions of Canada experience water-related problems, such floods, droughts, and water quality deterioration. Therefore, the field of water resources is one of the highest-priority fields with respect to climate change impacts and adaptation in Canada (Lemmen & Warren, 2004).

Canadian climate appears to be generally warmer and wetter during the last half of the 20th century. Gan (1992) found significant warming particularly in January, March, April, May and June over the past 40 years in western Canada. Whitfield & Cannon (2000) compared meteorological data for Canada from two different decades, and found the more recent decade to be generally warmer. According to Zhang *et al.* (2000), the annual average air temperature exhibits an increasing trend in southwestern Canada and a decreasing trend in the northeastern Canada. With respect to precipitation, Zhang *et al.* (2001) found that the annual precipitation totals have changed by –10% to 35%. These authors identified significant decreasing trends in winter precipitation and in the proportion of spring precipitation falling as snow in southeastern Canada. Whitfield (2001) found significant precipitation decreases during the autumn and significant increases during the winter and spring in British Columbia and the territory of Yukon.

The frequency and magnitude of extreme precipitation is also projected to change in Canada. For example, Zwiers & Kharin (1998) estimated that under a $2\times CO_2$ scenario, the return period of extreme precipitation would be shortened by half in northern Canada. Boer *et al.* (2000) found the largest rise in the average air temperature in the Interior Plains. Also, Stone *et al.* (2000) projected an overall increase in heavy rainfall frequency in the province of British Columbia (BC).

Changes in Canadian river flow have been extensively studied (cf. Westmacott & Burn, 1997; Coulson, 1997; Zhang *et al.*, 2001; Burn & Hag Elnur, 2002; Whitfield *et al.*, 2002; Morrison *et al.*, 2002; Yue *et al.*, 2003, 2004; Burn *et al.*, 2004) and correspond broadly to the regional changes identified in Canadian climate. Westmacott & Burn (1997) found decreases in the average monthly river flow that can be related to changes in the air temperature in the period of May–September in the Canadian Prairies. Coulson (1997) found an average annual runoff increase of 19–28%, corresponding to a precipitation increase of 7–18% in northern British Columbia. Zhang *et al.* (2001) calculated trends for 11 hydrometric variables for various Canadian catchments and found generally decreasing trends in river flow volumes, particularly in August and September. They observed significant increases in March and April river flows. Burn & Hag Elnur (2002) investigated 18 hydrological variables from a network of 248 catchments across Canada, and found large geographical differences in the results, implying that the impacts will not be spatially uniform. Whitfield *et al.* (2002) indicated that rainfall-driven streams in Georgia Basin, British Columbia demonstrate increased winter flows. Morrison *et al.* (2002) explored the climate change impacts in the Fraser River, British Columbia. They found a modest increase in the average river flow (5%) but a decrease in the average river peak flow (18%).

Snowmelt is an important source of river runoff as well as a significant flood-producing mechanism in many parts of Canada. Burn (1994) identified a trend in the timing of peak snowmelt runoff events for catchments in west-central Canada with more recent events occurring earlier in the year. According to Brugman *et al.* (1997), the late summer snowline in southern BC may rise up to 300 m with a doubling of CO_2. Under such conditions, less than 30% of the glacial surface will be covered by snow, leading to enhanced ice melt. Mote (2003) found clear evidence of warming-induced snow-pack declines, around 30% since 1950, particularly in spring and at lower elevations in the Georgia Basin–Puget Sound region. Fleming & Clarke (2003) found annual river flow volumes increasing in glacier-fed rivers of southwestern Yukon and northwestern BC. Whitfield (2001) examined impacts of recent warming on the hydrological regime in south-central British Columbia. This author found that snowmelt–induced runoff starts earlier, late summer and early autumn flows are lower, and early winter flows tend to be higher. Also, Leith & Whitfield (1998), Cannon & Whitfield (2001), and Cunderlik & Burn (2002) found shifts in the timing of the freshet and recession periods in British Columbia, causing significantly decreasing trends at the beginning of summer (reduced snowmelt-induced flows), and increasing trends in spring (increased snowmelt-induced flows). Earlier occurrence of spring river peak flow as a consequence of a warming trend in spring air temperatures was also identified in the Hudson Bay region (Gagnon & Gough, 2002) and in mainland Nunavut (Spence, 2002).

Changes in the frequency of hydroclimatic extremes may be one of the most significant consequences of climate change (Beven, 1993; Jones, 1999). Kite (1993) showed an increase in maximum river flows that is consistent with the projected increase in extreme rainfall events in the Rocky Mountains of British Columbia. Loukas & Quick (1999) and Loukas *et al*. (2002) investigated the potential impacts of the future climate change on the causes of flood flows for two mountainous watersheds located in two different climatic regions of British Columbia. The results showed that the overall flood magnitude and frequency of occurrence would increase in the coastal basin, and decrease in the interior basin. Roy *et al*. (2001) investigated the impact of climate change on summer and autumn flooding on the Chateauguay River basin in Québec. They indicated potentially very serious increases in the volume of runoff, maximum discharge and water level under future climate change scenarios. Whitfield *et al*. (2003) found increasing frequency of floods in all analysed watersheds in Georgia Basin, British Columbia. Weston *et al*. (2003) found peak annual flows of the Englishman River on the Vancouver Island to be 8% larger by 2020, 14% by 2050 and 17% larger by 2080. Cunderlik & Burn (2004) identified increasing maximum flows in spring, and decreasing autumn maximum flows in British Columbia.

Droughts are also projected to become more severe according to most scenarios of future climate in Canada. Yulianti & Burn (1998) investigated the impact of air temperature change on low-flow conditions for 77 rivers in the Canadian Prairies and found that low flows have a decreasing tendency. Hengeveld (2000) projected more frequent occurrence of droughts. The drought of 2001 affected Canada from coast to coast, with significant economic and social impacts. Many areas experienced the lowest summer precipitation in historic record (Lemmen & Warren, 2004). In 2001, the level of the Great Lakes reached its lowest point in more than 30 years (Mitchell, 2002). Significant trends toward longer frost-free periods could increase drought occurrence, since a longer ice-free season for lakes and rivers increases the potential for open-water evaporation (Environment Canada, 2004).

Most hydrological studies use the so-called impact approach to assess the potential consequences of climate change to the basin river runoff. The impact approach has three steps (IPCC, 2001): (1) calibration and verification of a hydrological model using observed hydroclimatic data; (2) derivation of climate change scenarios by perturbing observed series with increments deduced from GCM simulations; and (3) run of the model under new climate conditions and analysing impacts by comparing the results with the baseline simulation. The main objective of this study is to use the impact approach to assess the potential consequences of a changed climate to the timing and magnitude of hydrological extremes in a densely populated and urbanized river basin in southwestern Ontario, Canada. An ensemble of future climate scenarios is developed by means of a weather generating algorithm, linked with GCM outputs. The climate scenarios are then used as input into a semi-distributed hydrological model of the study river basin. The weather generating and hydrological models are described in the next section. This is followed by a description of the case study results. The last section summarizes the results.

METHODOLOGY

K-nn weather generator

The spatial and temporal scales of contemporary GCM outputs are inadequate for modelling hydrological processes at local, river basin scales. The outputs are therefore downscaled into scales more appropriate for hydrological modelling. Downscaling techniques include statistical downscaling (Conway & Jones, 1998; Sailor *et al*., 2000), dynamic downscaling (Jones *et al*., 1995; Frei *et al*., 1998; Giorgi *et al*., 1998; Jones & Reid, 2001), and downscaling based on stochastic weather generators (Schnur & Lettenmaier, 1998; Wilks, 1999; Yates *et al*., 2003).

Stochastic weather generators (WGs) have been increasingly applied to climate change impact studies. Weather generators allow for the creation of an ensemble of climate scenarios that can be used in integrated assessment studies. They can be parametric (Katz, 1977; Buishand, 1978; Woolhiser & Roldan, 1982; Richardson & Wright, 1984; Wilks, 1992; Semenov & Barrow, 1997; Parlange & Katz, 2000) and nonparametric (Young, 1994; Lall *et al*., 1996; Buishand & Brandsma, 2001; Yates *et al*., 2003). Nonparametric WGs can overcome most limitations of parametric WGs, such as probability distribution assumptions, reproduction of spatio-temporal data dependence, inability to capture non Gaussian data features, large number of parameters, and site-specific assumptions. Weather generators based on kernel multi-variate probability density estimators and K-nearest-neighbour (K-nn) bootstrap resampling methods are presently at the forefront of nonparametric weather generating techniques (Lall & Sharma, 1996; Rajagopalan & Lall, 1999; Buishand & Brandsma, 2001).

The K-nn approach based on the work of Yates *et al*. (2003) and Sharif & Burn (2004) is adopted in this study. The K-nn nonparametric WG uses the Mahalanobis distance metric, which does not require explicit weighting and standardization of variables. The Mahalanobis distance weights the variables with their covariance, which attributes less weight to strongly correlated variables. The algorithm samples the daily weather at the stations within a region simultaneously, and thus preserves the correlation structure between the stations and among variables. The bootstrap resampling retains non-Gaussian features in the probability density functions of the model variables. The approach can also simulate inter-annual and intra-annual climate variability. Technical details of the approach are described in Yates *et al*. (2003).

The K-nn algorithm can be used to perform strategic resampling to derive new daily weather data with altered mean and/or variability. Strategic resampling generates synthetic weather sequences from the historical record based on prescribed conditioning criteria. For a given climatic variable, regional periodical deviations are calculated for each year and for each period. For a particular period, a ranked list of years can be then generated by sorting the years according to the magnitude of deviations of that period. An index is then assigned to each year in the ranked list based on the relative position of the year in the sorted list. Different years from the ranked list are finally selected by means of a general integer function of the form (Yates *et al*., 2003):

$$f_{\Delta t}^{i} = \mathrm{Int}[n(1 - x^{\gamma_{\Delta t}^{i}})] + 1 \tag{1}$$

where $f_{\Delta t}^{i}$ is the index corresponding to year i and period Δt, x is a N (0,1) random number, and $\gamma_{\Delta t}^{i}$ is the shape parameter that can be adjusted to bias certain years over others. For example, if the years in the ranked list are arranged such that the coldest year has an index of 1 and the warmest year has an index of n, then $\gamma_{\Delta t}^{i} > 1$ would create bias towards the selection of warmer years, $\gamma_{\Delta t}^{i} < 1$ would create bias towards selection of colder years, and $\gamma_{\Delta t}^{i} = 1$ would lead to no bias.

HEC-HMS hydrological model

During the last decade a wide range of precipitation–runoff models has been used to assess the impact of global climate change on river basin hydrological processes (for a review, see e.g. Boorman & Sefton, 1997; Bronstert *et al*., 2002; and Bronstert, 2004). The US Army Corps of Engineers (USACE) Hydrologic Engineering Center Hydrologic Modeling System (HEC-HMS**)** has been applied successfully in many geographical areas for solving a variety of hydrological problems (Yu *et al*., 1999, 2002; Moges *et al*., 2003; Fleming & Neary, 2004; Cunderlik & Simonovic, 2004). The HMS model is a precipitation–runoff model that includes a large set of methods to simulate river basin, channel and water control structures.

The continuous simulation version of the HMS model used in this study includes seven components that describe the main hydroclimatic processes in the river basin. The meteorological component is the first computational element by means of which precipitation input is spatially (interpolation, extrapolation) and temporally (interpolation) distributed over the river basin. The spatio-temporal precipitation distribution is accomplished by the inverse-distance interpolation method (IDM). The IDM algorithm computes hyetographs for all selected locations in the river basin. A quadrant system is drawn centred on a given location. A weight for the closest rainfall gauge, that does not have missing data, is computed in each quadrant as the inverse, squared distance between the gauge and the location. The closest rainfall gauge in each quadrant is determined separately for each time step. When the closest gauge has missing data, then the next closest gauge in a quadrant is automatically used (USACE, 2000a).

The present version (v2.2.2) of HEC-HMS does not account for snow accumulation and melt processes. Since these processes are important flood-producing mechanisms in the study area, an external snow model was developed and linked with

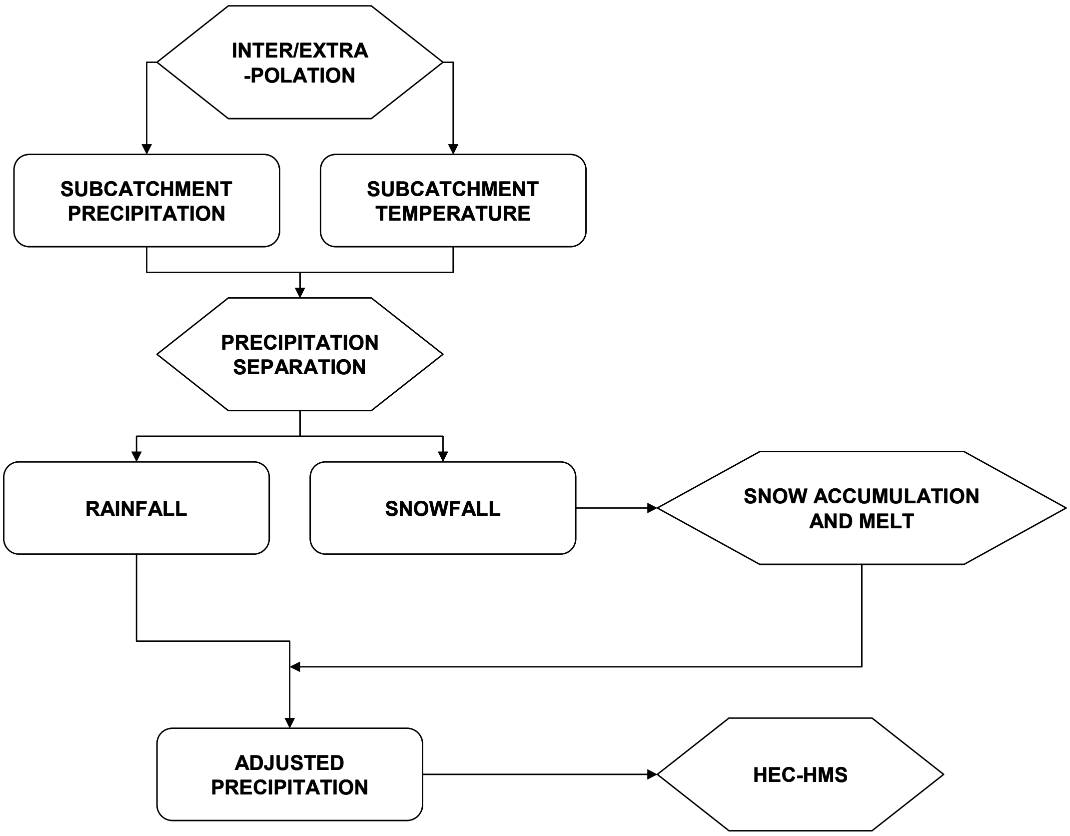

Fig. 1 Flow chart of the snow model.

the HEC-HMS. The snow model separates spatially- and temporally-distributed precipitation into liquid and solid forms, and simulates solid precipitation accumulation and melt. The algorithm of the snow model is based on a degree-day method. Degree-day models are common in snowmelt modelling due to wide availability of air temperature data, good model performance, and computational simplicity. In fact, most operational runoff models rely on degree-day methods for snowmelt modelling (Hock, 2003). Figure 1 illustrates the algorithm of the snow model. The precipitation for the time interval Δt is separated into solid (snowfall) or liquid (rainfall) form based on the average air temperature for the time interval Δt. The solid precipitation is then subject to the snow accumulation and melt algorithm. At each time interval Δt, the melted portion of snow, if any, is added to the liquid precipitation amount. The adjusted precipitation is then used as an input into the HEC-HMS model.

Precipitation adjusted by the snow component falls on pervious and impervious surfaces of the river basin. Precipitation from the pervious surface is subject to losses (interception, infiltration and evapotranspiration) modelled by the precipitation loss component. The 5-layer soil-moisture accounting (SMA) model is used to estimate and subtract the losses from precipitation. The SMA model is based on the Precipitation–Runoff Modeling System of Leavesley *et al*. (1983), and can be used for simulating long sequences of wet and dry weather periods. There are four different types of storage in the SMA model: canopy-interception storage, surface-depression storage, soil-profile storage, and groundwater storage (the model can include either one or two groundwater layers). The movement of water into, out of, and between the storage layers is administered by precipitation (input into the SMA system), evapotranspiration (output), infiltration (movement of water from surface storage to soil storage), percolation (from soil storage to groundwater storage 1), deep percolation (from groundwater storage 1 to groundwater storage 2), surface runoff (output), and groundwater flow (output). For computational details of the SMA model, see USACE (2000a). Precipitation from the impervious surface runs off with no losses, and contributes to direct runoff.

The output from the precipitation loss component contributes to direct runoff and to groundwater flow in aquifers. The Clark unit hydrograph (Clark, 1945) is used for modelling direct runoff. In the Clark method, overland flow translation is based on a synthetic time–area histogram and the time of concentration, *Tc*. Runoff attenuation is modelled with a linear reservoir. The groundwater flow is transformed into baseflow by a linear reservoir baseflow model. In this model, outflows from SMA groundwater layers are routed by a system of baseflow linear reservoirs.

Both overland flow and baseflow enter the river channel. The translation and attenuation of water flow in the river channel is simulated by the modified Puls method (USACE, 2000a). This method can simulate backwater effects (e.g. caused by dams), can take into account flood-plain storage, and can be applied to a broad range of channel slopes. The modified Puls method is based on a finite difference approximation of the continuity equation, coupled with an empirical representation of the momentum equation. The effect of hydraulic facilities (reservoirs, detention basins) and natural depressions (lakes, ponds, wetlands) is reproduced by the reservoir component of the model. Outflow from the reservoir is computed with the level-pool routing model. The model solves recursively one-dimensional approximation of the continuity equation.

Hydrological variables

A number of hydrological measures exists to describe the statistical properties of extreme hydrological events. In terms of high flows, the following five measures are considered in this study:

– Absolute maximum daily river flow, *AMAX*, defined as the highest daily river flow recorded during the observation/simulation period.
– Average annual maximum daily river flow, *MAX*, defined as the average value of a time series consisting of annual maximum daily river flows. *MAX* time series are often used in flood frequency analysis if instantaneous river flow series are not available. The annual maximum flows are determined here for water years.
– The timing, *DMAX*, and regularity, *RMAX*, of annual maximum daily flows, described in terms of directional statistics (see e.g. Fisher, 1993). The variable *DMAX* represents a measure of the average day of occurrence of annual maximum river flow events. The variable *RMAX* provides a dimensionless measure of the regularity of annual maximum river flow occurrences. A value of unity indicates that all annual maximum river flow events in the sample occurred on exactly the same day of the year, while a value closer to zero indicates that there is greater variability in the date of occurrence of annual maximum river flow events. A

methodology for calculating *DMAX* and *RMAX* variables can be found in Bayliss & Jones (1993).

- Coefficient of variation of annual maximum daily flows, *CMAX*, a dimensionless measure of the variability of annual maximum river flow magnitudes.
- The number of high flows, *NHF*, i.e. flows greater than $\overline{X} + 3 \times S_x$, where $\overline{X}$ and S_x are the average and standard deviation of daily river flow series.

In terms of low flows, the following six hydrological measures are used in this study:

- Absolute minimum daily river flow, *AMIN*, defined as the lowest daily river flow recorded during the observation/simulation period.
- Average annual minimum daily river flow, *MIN*, defined as the average value of a time series consisting of annual minimum daily flows.
- The timing, *DMIN*, and regularity, *RMIN*, of annual minimum daily flows, calculated according to the same procedure as *DMAX* and *RMAX* measures.
- Coefficient of variation of annual minimum daily flows, *CMIN*, describing the variability of annual minimum river flow magnitudes.
- Dry weather flow, *MIN7*, defined as the average annual 7-day minimum river flow.
- The 90th percentile flow, *Q*90, defined as the river flow which is equalled or exceeded for 90% of the period of record. The *Q*90 measure is determined from a flow duration curve representing the relationship between the magnitude and frequency of daily river flows.

A detailed review of low-flow hydrology, low-flow characteristics and their applications can be found in Smakhtin (2001).

Model calibration and performance

A systematic manual calibration was chosen for setting up the hydrological model. The calibration relies primarily on the measured and estimated values of the model parameters available from the study area. This ensures that a physically meaningful set of initial parameter values is used for the calibration. The calibration parameter thresholds are defined as initial parameter value ±75%. The performance of the hydrological model at the end of each calibration trial is assessed by the following four statistical measures:

1. Average absolute error in peak magnitudes (*AEPM*):

$$AEPM = \frac{1}{n}\sum_{i=1}^{n}\left|\frac{Q_O^P - Q_M^P}{Q_O^P}\right| \times 100\,[\%] \qquad (2)$$

where n is the number of annual maximum observed, Q_O^P, and modelled, Q_M^P, peak flows.

2. Error in runoff volume (*EV*):

$$EV = \frac{V_O - V_M}{V_O} \times 100\,[\%] \qquad (3)$$

where V_O is the observed, and V_M modelled total runoff volume.

3. Root mean squared error (*RMSE*):

$$RMSE = \sqrt{\frac{1}{N}\sum_{t=1}^{N}\left(\frac{O_{M\,t} - O_{O\,t}}{O_{O\,t}}\right)^2} \times 100\,[\%] \tag{4}$$

where Q_{Ot} and Q_{Mt} are, respectively, the observed and modelled river flow at time t, and N is the number of hydrograph ordinates.

4. Nash & Sutcliffe (1970) efficiency criterion (E):

$$E = 1 - \frac{\sum_{t=1}^{N}\left(Q_{Ot} - Q_{Mt}\right)^2}{\sum_{t=1}^{N}\left(Q_{Ot} - \overline{Q_O}\right)^2}\,[\text{-}] \tag{5}$$

where $\overline{Q_O}$ is the average observed river flow for the simulation period.

Each calibration trial is assessed according to the above described criteria. If the performance of the model is acceptable, the calibration process is completed, otherwise the initial calibration parameters are altered and the process repeated. The calibration of a semi-distributed model starts with hydrometric stations that represent outlets of single sub-basins. Once these stations are calibrated, hydrometric stations with more than one contributing sub-basin follow. At this stage the parameters of ungauged contributing sub-basins are also estimated. In the final stage, individually calibrated sub-basins are linked into one model and the calibration is finalized.

Klemeš (1986) discusses different validation tests used in hydrological modelling. Among them, the split sample test is the most commonly used technique. It involves dividing the available measured time series into two sets. The parameters of the model are calibrated using one set. The validity of the model is then tested by running the calibrated model using the input data from the second set, and comparing the resulting projections for the output variables against the measured values. This approach is also used in this project. Only the parameters describing river basin initial conditions (initial storage of the canopy, surface, soil and groundwater layers) are changed over time during the verification of the model. All other parameters of the model are kept constant while the model is tested using the separate validation data set.

STUDY AREA

The Upper Thames River Basin (UTRB) is located in southwestern Ontario, Canada. The UTRB has a drainage area of 3450 km^2 and outlets to the Lower Thames River, which is a tributary to Lake St. Clair. The population of the UTRB is 460 000. The main urban centre in the UTRB is the city of London, which is designated a growth centre in the province of Ontario. Urban growth is contrasted by intensive farming in the basin.

The Thames River corridors, located in a highly developed part of southwestern Ontario are vulnerable from both urban and rural land-use pressures. Despite these pressures, the Thames remains one of the most biologically diverse rivers in Canada. In fact, the Thames River system is declared a Canadian Heritage River based on its rich cultural heritage and diverse recreational opportunities.

Floods and droughts represent the main hydrological hazards in the UTRB. Snowmelt is a major flood-producing factor in the basin, generating flood events most frequently in March. Intensive flood-producing storms are most frequent in August. Periods of low flows usually occur during the summer, and the risk of droughts is highest in the months of July and August. Three main reservoirs in the UTRB: Fanshawe, Wildwood, and Pittock, assist in flood management efforts and river flow augmentation during the drier summer months.

The area of UTRB is divided into 32 smaller sub-basins representing fine spatial resolution for semi-distributed hydrological modelling by means of the US Army Corps of Engineers HEC-GeoHMS software (USACE, 2000b). The meteorological component interpolates climatic input data into 32 different UTRB locations, one for each sub-basin defined as the sub-basin's centroid. The HMS model parameters are assumed to be uniform within each sub-basin. The channel network is derived from a digital elevation model (DEM). The DEM is conditioned by removing sinks and pits. The delineated sub-basins, evapotranspiration zones, and 15 hydrometric and 15 precipitation observation stations available in the study area are depicted in Fig. 2.

RESULTS

Model performance

The 9-year long observation period from November 1979 to October 1988 is selected for the calibration, and the 9-year long observation period from November 1988 to October 1997 for the verification of the hydrological model. Both periods have the highest spatio-temporal data density in the UTRB, as well as hydrological variability representative of the whole available historical records. An 8-year long calibration period or longer should, according to Yapo *et al*. (1996) assure that the results will be insensitive to the period selected. Also Xu & Vandewiele (1994) found that a data length of around 10 years is sufficient for a reliable calibration of hydrological models.

In accordance with the temporal resolution of the K-nn weather generator, a daily time step is chosen to represent the computational time interval for the hydrological model. Preliminary results obtained from the hydrological model calibration systematically underestimated winter and spring river flows and overestimated summer and autumn river flows. This error is likely associated with the discrepancy between the nonlinear precipitation–runoff response in the UTRB and the linear structure of the SMA model. A semi-annual parameterization approach is therefore applied, in which separate parameter sets are established for summer and winter seasons. The summer season is defined from 1 May to 31 October, and the winter season from 1 November (the beginning of a water year in the study area) to 30 April. A semi-annual model is recommended also by Fleming & Neary (2004), who showed that the performance of a semi-annual HEC-HMS model is better than the annual, single-parameter set model.

The semi-annual approach applied in this study separates parameters that can take different values in summer and winter seasons from the parameters that are assumed

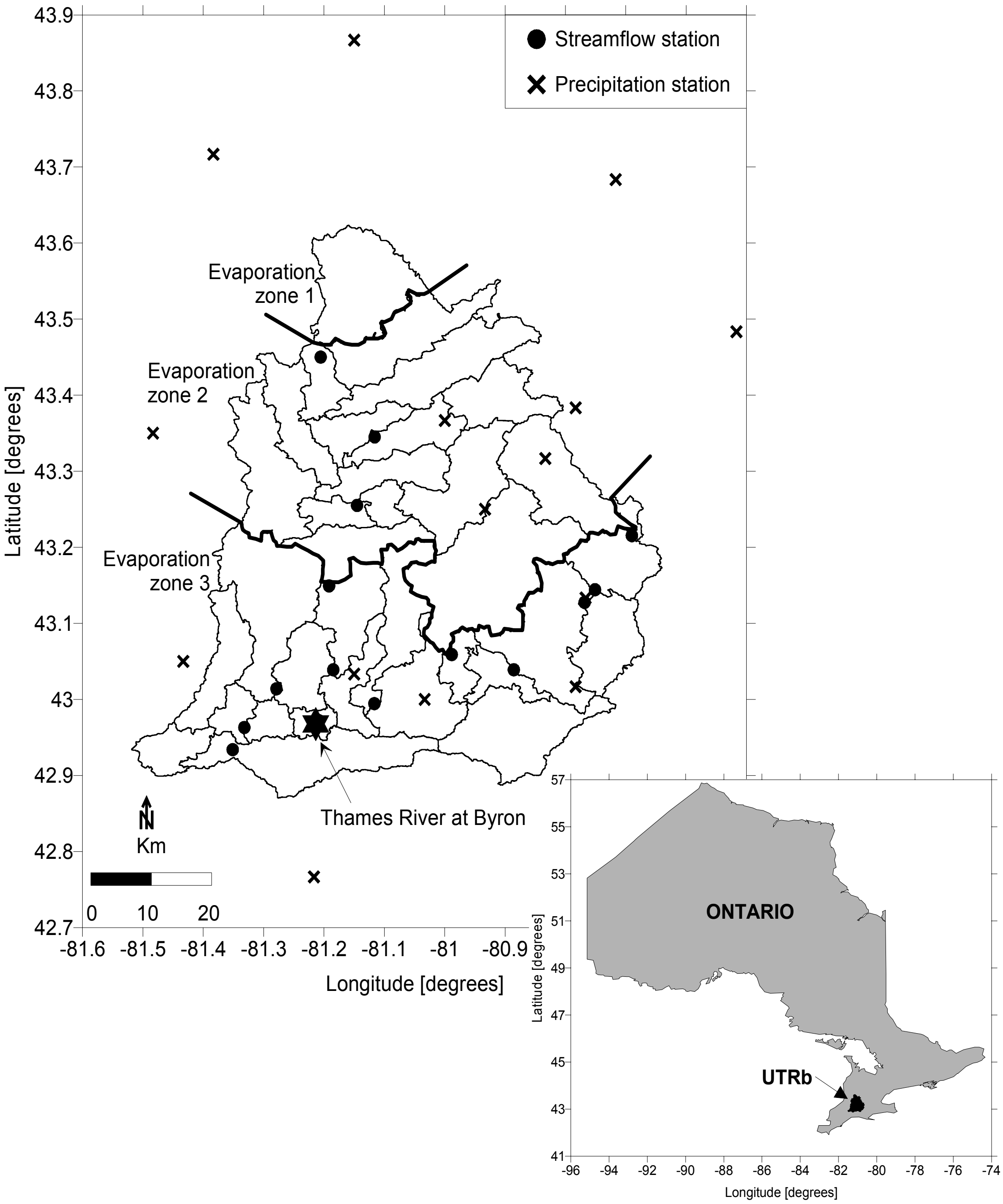

Fig. 2 The division of the Upper Thames River basin into 32 sub-basins, evapotranspiration zones, and the locations of streamflow and precipitation stations used in the study.

seasonally invariant. Apart from the parameters describing river basin initial conditions, the SMA surface storage capacity and the maximum soil infiltration rate are considered as seasonally dependent parameters. The rationale behind the seasonal alteration of these parameters is that in winter, precipitation is mostly accumulated on the surface, which alters the attenuation of water represented by the surface storage parameter. Further, the dominant solid precipitation and changed physical properties of soils (such as lower hydraulic conductivity due to frozen water in soil pores) reduce winter soil infiltration rates. The remaining model parameters are deemed constant in both summer and winter models.

Figures 3 and 4 compare the observed and modelled daily river flow hydrographs for the Thames River at Byron (Water Survey of Canada ID 02GE002, see Fig. 2 for site location) for two time windows selected from the calibration and verification periods. The external snow model adequately reproduces the snow accumulation and melt processes. In particular, the temporal occurrence of spring snowmelt-generated peaks is well captured by the model. Secondly, there is no systematic bias in the winter season peaks or summer season peaks in the seasonal version of the model. Also, the performance of the model in the simulation of dry periods of low flows is good. The model performance improves with the increase of basin area and spatial detail.

Table 1 summarizes the statistical performance measures defined by equations (2)–(5) obtained from the data used for the calibration and verification of the hydrological model. The average error in peak magnitudes, *AEPM*, is rather large for both calibration and verification periods, but the model is almost unbiased in terms of peak magnitudes. Indeed, when the *AEPM* is replaced by a relative measure (sign of the peak errors in equation (2) is taken into account), the average relative error is only 3.4% for the calibration period, and 4.7% for the verification period. The model slightly underestimates total river flow volumes by 8–9%. In terms of *RMSE* and *E* measures, the model performs similarly with the data from the calibration and verification periods.

The performance of the model is assessed also by evaluating the model ability to reproduce the extreme hydrological measures described in the previous section. Table 2 compares the hydrological variables obtained from the observed and modelled daily river flow series from the 30-year long baseline period 1971–2000. The variables describing high flows are captured well by the model, with relative differences between the modelled and observed measures only up to around 5%. However, the low flow hydrological variables are underestimated by the model. The average annual minimum river flow values are 25% lower than the observed values, the variability of annual minimum river flow occurrence is underestimated by –38% (this actually means that the variability of modelled low flows is higher, see definition of the *RMIN* measure), dry weather flow by –20%, and the modelled absolute minimum daily river flow is 1.58 $m^3 s^{-1}$ in contrast to the observed 3.5 $m^3 s^{-1}$. The deviation of the 90th percentile flow as well as the timing and variation of annual minimum river flows from the baseline values are within ±5%.

The above described limitations of the model are considered in the subsequent evaluation of the model outputs obtained from future climate scenarios. Also, the model is calibrated and verified on spatially and temporally interpolated precipitation. The spatial and temporal distribution of the interpolated precipitation may not always correspond well to the true precipitation distribution. The interpolated precipitation pattern is thus reflected in the calibrated model parameters. The calibrated model parameters are deemed to be climate change invariant based on the assumption that the change in processes represented by the parameters will be small in comparison with the changes affecting the climatic conditions (Drogue *et al.*, 2004).

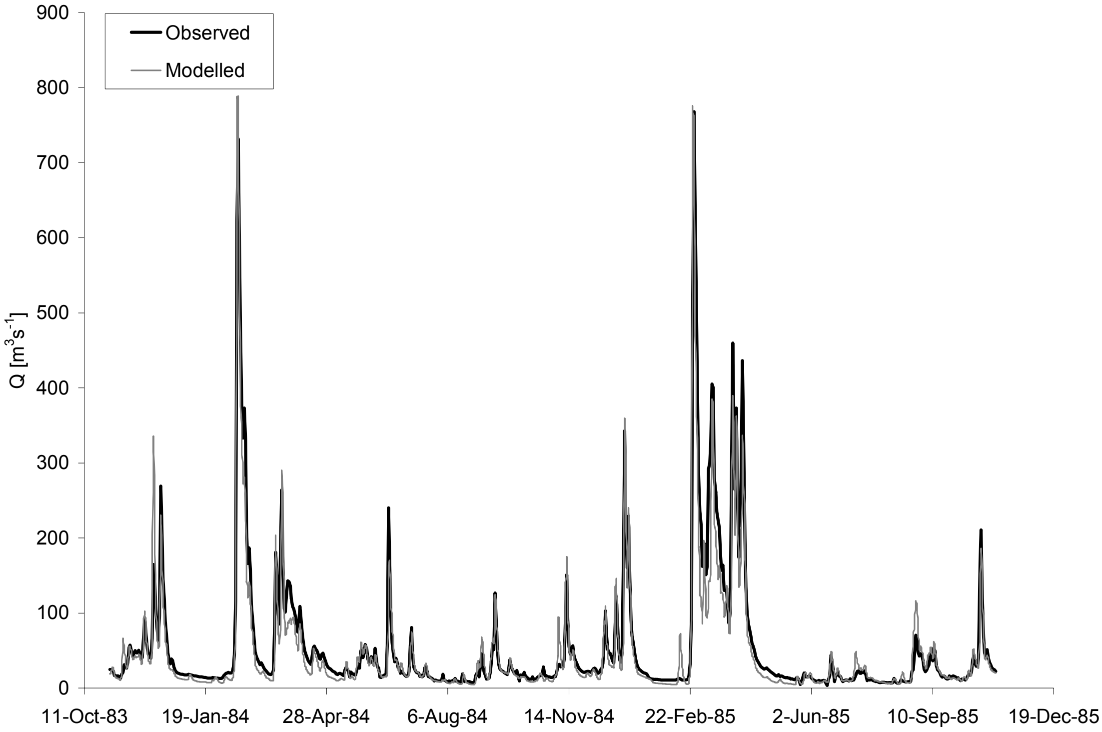

Fig. 3 Observed and modelled hydrographs of the Thames River at Byron for the calibration period (1 November 1983–31 October 1985 sequence shown).

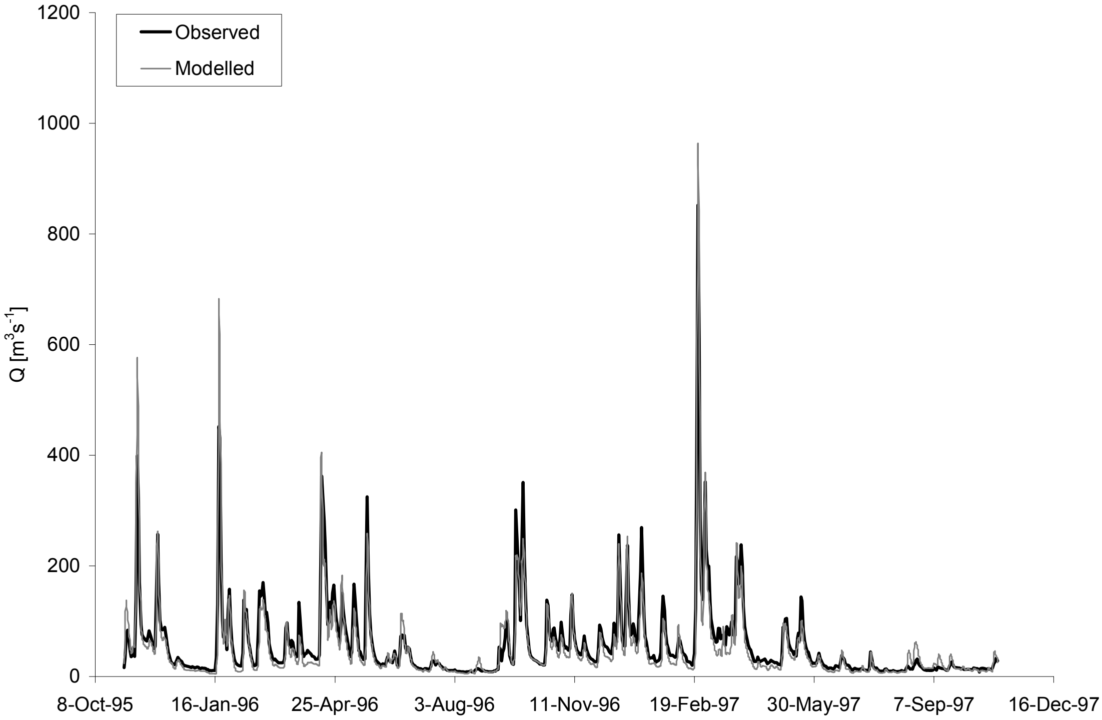

Fig. 4 Observed and modelled hydrographs of the Thames River at Byron for the verification period (1 November 1995–31 October 1997 sequence shown).

Table 1 Performance of the HEC-HMS hydrological model.

Measure	Period	*AEPM* (%)	*EV* (%)	*RMSE* (%)	*E* (-)
Calibration	Nov. 1979–Oct. 1988	18.261	8.827	44.826	0.891
Verification	Nov. 1988–Oct. 1997	14.875	7.607	46.488	0.863

AEPM: average absolute error in peak magnitudes; *EV*: error in runoff volume; *RMSE*: root mean squared error; *E*: Nash & Sutcliffe efficiency criterion.

Table 2 Hydrological variables determined from the observed and modelled 30-year baseline period 1971–2000.

Variable	Observed	Modelled	Difference (%)
AMAX ($m^3 s^{-1}$)	852.000	883.150	3.527
MAX ($m^3 s^{-1}$)	522.526	546.820	4.443
DMAX (°)	52.013	50.700	–2.590
RMAX (-)	0.714	0.671	–6.408
CMAX (-)	0.351	0.357	1.681
NHF (#)	169	176	3.977
AMIN ($m^3 s^{-1}$)	3.500	1.580	–121.519
MIN ($m^3 s^{-1}$)	6.776	5.428	–24.834
DMIN (°)	221.317	216.872	–2.050
RMIN (-)	0.685	0.496	–38.105
CMIN (-)	0.395	0.378	–4.497
MIN7 ($m^3 s^{-1}$)	7.974	6.651	–19.892
*Q*90 ($m^3 s^{-1}$)	8.800	8.804	0.045

AMAX: absolute maximum daily flow; *MAX*: average annual maximum daily flow; *DMAX*: timing of annual maximum daily flows; *RMAX*: regularity of annual maximum daily flows; *CMAX*: coefficient of variation of annual maximum daily flows; *NHF*: number of high flows; *AMIN*: absolute minimum daily flow; *MIN*: average annual minimum daily flow; *DMIN*: timing of annual minimum daily flows; *RMIN*: regularity of annual minimum daily flows; *CMIN*: coefficient of variation of annual minimum daily flows; *MIN7*: dry weather flow; *Q*90: 90th percentile flow.

Sensitivity analysis

A local sensitivity analysis is adopted for evaluating the parameters of the hydrological model. The calibrated model parameters are considered as the baseline/nominal parameter set. The model is run repeatedly with the baseline value for each parameter multiplied, in turn, by 0.8 and 1.2, while keeping all other parameters constant at their nominal starting values. The hydrographs resulting from the scenarios of adjusted model parameters are then compared with the baseline model hydrograph. The sensitivity procedure is applied to the Middle Thames River at Thamesford (ID 02GD004), using the precipitation data from the time period 1 November 1983–31 October 1985. The Middle Thames River at Thamesford is centrally located in the UTRB, relatively pristine, and representative in terms of the UTRB hydroclimatic regime. The results are summarized in Table 3. With respect to flood magnitudes *AEPM* and *EV*, the Clark's storage coefficient *St* and the parameters describing physical properties of the soil (infiltration rate *If* and soil layer storage *Us* and *Ts*) are the parameters that have the greatest impact on peak hydrographs generated by the continuous model. In terms of the peak volume, the continuous model is most sensitive to the three SMA groundwater layer parameters (*Gs*, *Gp*, and *Gc*). The SMA groundwater parameters in combination with the baseflow parameters are also impor-

Table 3 Sensitivity analysis of the HEC-HMS hydrological model for the Middle Thames River at Thamesford.

Parameter	Parameter change (%)		Parameter	Parameter change (%)	
Tc	–20	+20	*Us*	–20	+20
AEPM (%)	1.167	2.708	*AEPM* (%)	5.161	3.466
EV (%)	0.001	0.001	*EV* (%)	8.513	4.194
RRMSE (%)	3.031	3.054	*RRMSE* (%)	18.093	9.663
E (-)	0.898	0.898	*E* (-)	0.894	0.899
St	–20	+20	*Ts*	–20	+20
AEPM (%)	8.853	9.204	*AEPM* (%)	4.213	4.222
EV (%)	0.002	0.001	*EV* (%)	4.071	5.671
RRMSE (%)	8.844	5.385	*RRMSE* (%)	11.630	19.884
E (-)	0.897	0.898	*E* (-)	0.898	0.898
Bs	–20	+20	*Sp*	–20	+20
AEPM (%)	0.079	0.069	*AEPM* (%)	0.000	0.000
EV (%)	0.043	0.044	*EV* (%)	0.023	0.001
RRMSE (%)	26.967	16.634	*RRMSE* (%)	1.128	0.542
E (-)	0.997	0.997	*E* (-)	0.997	0.997
Br	–20	+20	*Gs*	–20	+20
AEPM (%)	0.059	0.099	*AEPM* (%)	0.494	0.412
EV (%)	0.052	0.919	*EV* (%)	8.697	6.839
RRMSE (%)	22.772	64.168	*RRMSE* (%)	24.569	28.689
E (-)	0.997	0.899	*E* (-)	0.997	0.884
Cs	–20	+20	*Gp*	–20	+20
AEPM (%)	0.000	0.000	*AEPM* (%)	0.548	0.445
EV (%)	0.114	0.078	*EV* (%)	8.433	7.416
RRMSE (%)	0.750	0.702	*RRMSE* (%)	29.307	33.466
E (-)	0.997	0.997	*E* (-)	0.883	0.886
Ss	–20	+20	*Gc*	–20	+20
AEPM (%)	0.538	0.544	*AEPM* (%)	0.655	0.494
EV (%)	1.458	1.337	*EV* (%)	8.091	7.105
RRMSE (%)	4.735	6.168	*RRMSE* (%)	12.119	12.254
E (-)	0.899	0.997	*E* (-)	0.899	0.899
			If	–20	+20
			AEPM (%)	2.607	2.488
			EV (%)	3.596	3.137
			RRMSE (%)	8.113	8.303
			E (-)	0.899	0.899

Tc: time of concentration; *St*: Clark's storage; *Bs*: baseflow storage; *Br*: number of baseflow reservoirs; *Cs*: canopy storage; *Ss*: surface storage; *Us*: soil storage; *Ts*: tension zone capacity; *Sp*: soil percolation rate; *Gs*: groundwater storage; *Gp*: groundwater percolation rate; *Gc*: groundwater storage coefficient; *If*: infiltration rate.

tant for simulating low flows, and for the overall goodness-of-fit of the continuous model.

Fleming & Neary (2004) performed a similar sensitivity analysis of a continuous HEC-HMS model of the Dale Hollow basin in Kentucky and Tennessee, USA. They concluded that the maximum infiltration rate, *If*, the soil storage depth, *Us*, and the tension zone depth, *Ts*, caused the most variation in simulated river flows when adjusted. While our results also include Clark's storage, the difference between the sensitivity results from Fleming & Neary (2004) and the results obtained in this study can be, apart from other factors, attributed to the limitation of a local sensitivity analysis, which depends on the actual combination of model parameters. For example, if the percolation rate between the soil layer and the first groundwater layer is high, then the model will be less sensitive to the parameters describing water content in the soil. On the other hand, if the deep percolation rate between the two groundwater layers is low, then the parameters of the first groundwater layer are likely to be the highly sensitive parameters of the SMA model, because the water will tend to retain in this layer longer than in the other layers of the system.

Hydrological extremes under future climate scenarios

The performance of the K-nn weather generator used in this study to replicate the present (baseline) climate in the UTRB and to simulate an ensemble of different future climate change scenarios is presented by Sharif & Burn (2004). According to these authors, the K-nn weather generator reproduces adequately all main statistical properties of observed daily data. Spatial and temporal dependencies in the data are also well preserved. The model reproduces well derived monthly data statistics, which further supports the good fit at the daily time step. For further details the reader is referred to Sharif & Burn (2004).

Strategic resampling is used to generate an ensemble of synthetic weather sequences from the historical record based on different scenarios prepared by the Canadian Climate Impacts Scenarios (CCIS) Group, using data from Canadian Centre for Climate modelling and analysis (CCCma). Here, only the results from two likely scenarios are presented, one with an increase of 1°C in the average daily temperature (IncTemp scenario hereafter) and the other one with an increase of 100 mm in the average annual precipitation (IncPrec scenario). Changes in the seasonal distribution of precipitation are not taken into account in this study. Evapotranspiration scenarios are not considered, and the UTRB basin is assumed unchanged in terms of natural and anthropogenic processes (soil properties, land use, stream channelling, drainage, etc.).

The hydrological model is run with a 30-year long sequence of climatic data produced by the weather generator for the two future climate scenarios. Figure 5 compares the hydrographs generated by the IncTemp and IncPrec scenarios with the baseline hydrograph for one simulated water year (one winter and one summer season). It can be seen that the IncTemp scenario produces more snowmelt induced peaks occurring earlier than the main snowmelt peak of the baseline data. The IncPrec scenario generates higher river flows in spring (March) and in autumn (October–November).

The characteristics of hydrological extremes introduced in the previous text are calculated from the model outputs obtained from the IncTemp and IncPrec scenarios, and compared to the characteristics obtained from the baseline data. The results are presented in Fig. 6. In terms of high flows, the IncTemp scenario projects almost a 50% decrease in the magnitude of the average annual maximum daily flows (*MAX*). In UTRB, annual maximum flows are generated dominantly by snowmelt. Under the IncTemp scenario, the period of snowmelt will occur earlier, will be longer, and melt

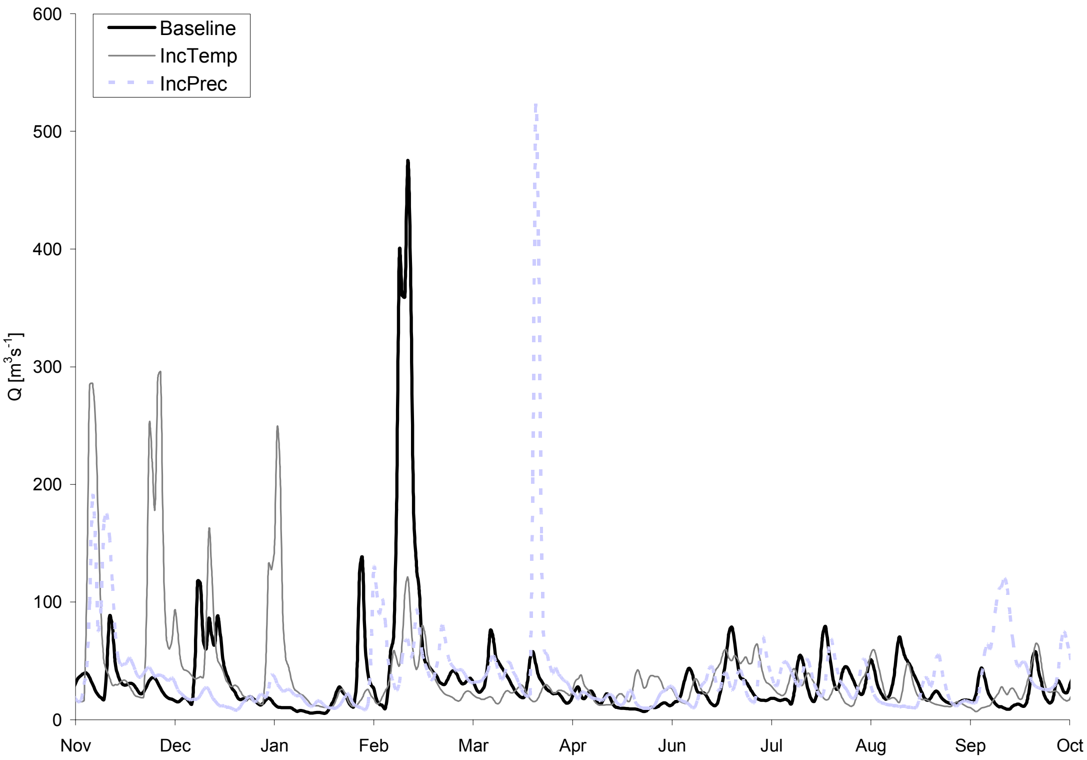

Fig. 5 A typical seasonal distribution of daily streamflow for baseline, increased temperature (IncTemp) and increased precipitation (IncPrec) scenarios.

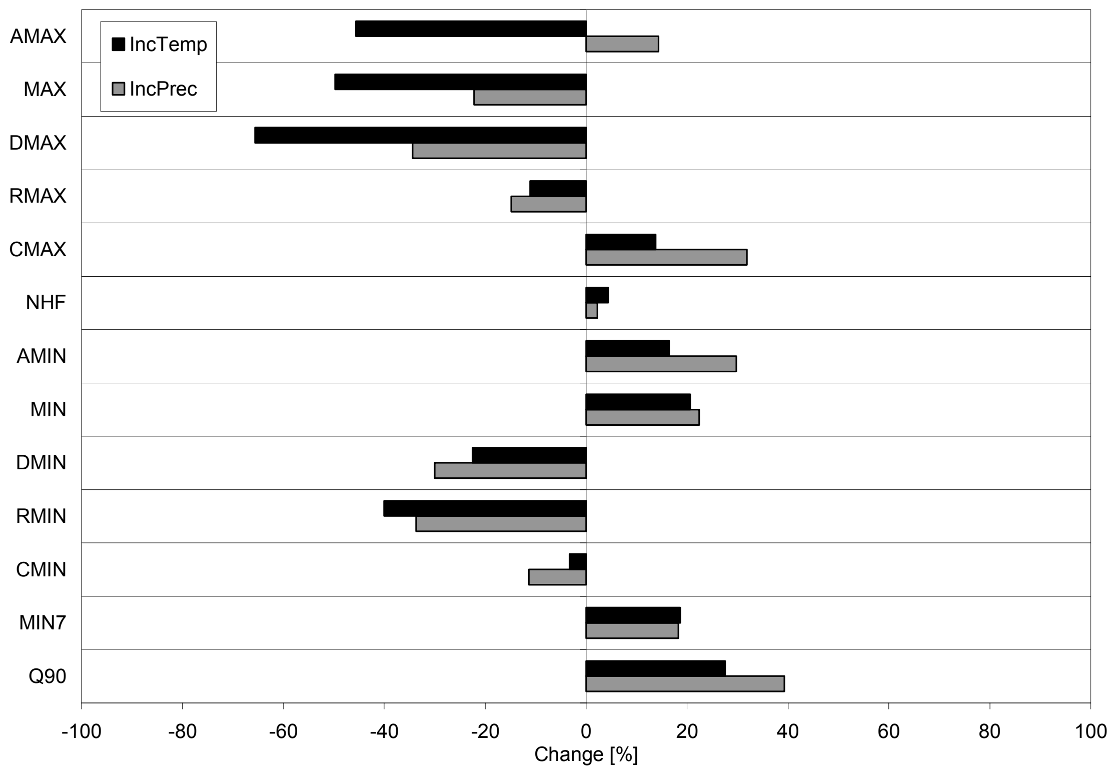

Fig. 6 Expected changes in the selected characteristics of hydrological extremes for the increased temperature (IncTemp) and increased precipitation (IncPrec) scenarios.

intensity more evenly distributed during this period. Figure 5 illustrates that the main snowmelt peak of the baseline run is replaced by several smaller snowmelt runoff peaks occurring earlier in the year under the IncTemp scenario. More snowmelt peaks also increases the number of high flows (*NHF*) by 4.5%. The decrease in the magnitude of the absolute maximum daily river flow, *AMAX* (–45%) for the increased temperature scenario is in good accordance with the decrease in the magnitude of the average annual maximum river flow.

The increased precipitation scenario generates a 20% decrease in the average annual maximum flow (*MAX*), and a 15% increase in the absolute maximum daily river flow (*AMAX*). This result can be explained by a positive correlation between temperature and precipitation during the winter months (November, December, January and February) that exists in the historical observation. Owing to this positive correlation, increases in precipitation in January and February are accompanied by increases in temperature values, which in turn generate snowmelt runoff earlier in the year. The IncPrec scenario produces higher rainfall events during the warm season, which generates some very high flood events, and consequently, a higher value of *AMAX*. More spring and autumn flood events increase the number of high flows (*NHF*) by 2.5%.

The shift of the snowmelt-produced *MAX* events toward the winter according to both scenarios is also reflected by the *DMAX* measure. The average occurrence of annual maximum flows at the end of February for the baseline data shifts to the beginning of January (IncTemp scenario) and to the beginning of February (IncPrec scenario). Also the *RMAX* measure drops by 15%, which means less regularity in the temporal occurrence of annual maximum river flow events. This is caused by the stretched melting period of snow as discussed earlier. On the other hand, the variability of annual maximum river flow magnitudes is expected to increase, around 15% for the IncTemp scenario and more than 30% for the IncPrec scenario, mainly due to higher rainfall amounts in the warm season.

In terms of extreme minimum flows, both scenarios project an increase in the absolute minimum daily river flow (*AMIN*) and the average annual minimum daily river flow (*MIN*). The *AMIN* value for the IncTemp (IncPrec) scenario is 15% (30%) higher then the baseline *AMIN* value. The increase in the *MIN* values is for both scenarios around 20%. Also the dry weather flow (*MIN7*) and the 90th percentile flow (*Q*90) characteristics of minimum flows, which reflect a wider distribution of daily flows, increases by 20–40%. The increase in minimum flows can be perhaps explained by the shift in the period of low flows from the beginning of August (baseline) toward the month of June, where the soil moisture storage is still affected by the wetter spring months. Indeed, the timing of annual minimum daily river flows (*DMIN*) decreases by 55–65 days compared to the *DMIN* value of the baseline data. Also the regularity of annual minimum daily river flows (*RMIN*) decreases by 35–40% for both scenarios, which means an increase in the duration of low flow (drought) period. However, the model performance in reproducing the *RMIN* characteristic of the historical data is rather poor (refer to Table 2), so the low flow regularity may not change as severely. In terms of the magnitude of annual low flows, the evaluated scenarios project a 5–15% decrease in the *CMIN* characteristic. In a summary, low flows may become less extreme, less variable in terms of magnitude, and more variable in terms of temporal occurrence.

The presented results correspond well with the results published in other studies focused on southwestern Ontario. For example FitzGibbon *et al*. (1993) estimated a 10% reduction in spring peak flows in the Grand River. De Loe *et al*. (2001) discuss some adaptation options for this region. Also Smith *et al*. (1998) projected for the region of southern Ontario more intense rainfall falling at an altered frequency of occurrence, and an increased number of mid winter melts as a consequence of milder winters.

CONCLUSIONS

Future hydraulic structures, flood-plain development, and water resource management will need to accommodate the uncertainties with respect to changes in flood and low-flow distributions. This study used the impact approach to assess the potential consequences of a changed climate to the timing and magnitude of hydrological extremes in a densely populated and urbanized river basin in southwestern Ontario, Canada. An ensemble of changed climate scenarios is developed by means of a weather generating algorithm, linked with GCM outputs. The climate scenarios are then transformed into basin runoff by a semi-distributed hydrological model of the study area.

The results of two likely scenarios of change in precipitation and temperature can be summarized as follows:

- Significant decrease in the magnitude of snowmelt-induced annual maximum daily flows can be expected due to longer snowmelt period stretched toward the winter, with more evenly distributed snowmelt intensities.
- The magnitude of rainfall-induced maximum flows will increase, particularly under the increasing precipitation scenarios.
- The occurrence of annual maximum flows under increased temperature and precipitation will be less regular than it is today.
- The variability of the magnitude of annual maximum flows is expected to increase.
- High river flow periods may be more frequent in the future.
- Increased temperature and precipitation scenarios project an increase in the magnitude of low flows.
- The variability of the magnitude of annual minimum flows is expected to decrease.
- The period of low flows (droughts) will stretch from August (baseline) toward June.
- The occurrence of annual minimum flows under increased temperature and precipitation will likely be less regular then it is today.

According to the two evaluated scenarios, climate change may have beneficial impacts on the distribution of hydrological extremes in the study area. The future regime of maximum flows in UTRB can be characterized as less extreme in terms of magnitude (dominant snowmelt-induced maximum flows will decrease), and more variable in terms of both occurrence and magnitude. Low flows may become less extreme, less variable in terms of magnitude, and more irregular in terms of occurrence.

With respect to the numerous uncertainties involved in climate change impact modelling, the sign of the changes projected by the model should be favoured over the absolute values. Only two extreme cases out of many possible scenarios are evaluated in this study. Different climate change scenarios will produce different river basin runoff responses. In order to quantify the uncertainty of the model outputs, a wider spectrum of scenarios needs to be evaluated, and the results of these ensembles assessed. The presented results are also regionally limited, since physical properties of the river basin (e.g. topography, drainage density, soil permeability), and other concurrent basin-specific changes (land use, reservoirs, stream channelling, drainage) play an important role in the impact modelling. The outcomes of the study will be translated into new hazard mitigation guidelines and vulnerability reduction strategies for improved flood prevention and robust water management under changing climatic conditions in the study area.

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