Nordic Hydrology, 23, 1992, 257-272

Sensitivity of Streamflow Simulation to Changes in Climatic Inputs

H. Y. F. Ng and J. Marsalek

National Water Res.Inst.,
Burlington, Ont., Canada L7R 4A6

Sensitivity of streamflow simulations to changes in temperature and precipitation were evaluated for a small basin ($A \equiv 53\ km^2$) characterized by low evapotranspiration and high conversion of precipitation into streamflow. Using a calibrated HSPF model, such sensitivities were studied for temperature increases up to 4° C and precipitation fluctuations of ± 10%. Temperature increases barely affected annual streamflow, led to larger and earlier winter runoff when precipitation was stored in the snowpack, and increased winter/spring streamflow peaks. Effects of precipitation fluctuations were more direct – annual and seasonal streamflow fluctuations were directly proportional to precipitation changes and monthly peaks increased about twice as much as precipitation.

Introduction

Growing atmospheric concentrations of radiatively-active gases contribute to global climate warming and the climatic changes which will affect the quantity and quality of the freshwater resources and may lead to major alterations of regional water systems (WMO 1987). Assuming a doubling of atmospheric CO_2 concentrations in the next 50 years, some earlier reports (U.S. Department of Energy 1980; IPCC 1990) estimated that temperatures in the Northern Hemisphere may rise between 3° and 5° C, and, depending on the latitude, precipitation may increase or decrease by as much as 15% (IPCC 1990).

Changes of that magnitude and speed may have far reaching impacts on the hydrological cycle in the form of new stresses on water availability, quality, and supply. Gross patterns of precipitation, evaporation and runoff are likely to change, but spatial and temporal details of such changes are unknown. While some impacts in the form of rising land-air and sea-surface temperatures have already been reported (IPCC 1990), others, in the form of increased annual discharges or lake levels, have not been detected (Kite 1989). Some of the predicted impacts on water resources can be mitigated in river basins with adequate storage and regulation (Klemes 1990).

The earlier studies of specific impacts of climate changes on water resources may be classified into two categories – studies using global climate models to predict impacts of climate change scenarios (U.S. Department of Energy 1980; Gleick 1987a; Cohen 1986; IPCC 1990) and studies using hydrologic simulation, with assumed hypothetical climatic inputs, to demonstrate changes in various components of the hydrologic cycle (Nemec and Schaake 1982; Nemec 1989; McCabe and Ayers 1989; Sanderson and Smith 1990; Thomsen 1990). While the global circulation models are invaluable tools for identifying climatic sensitivities and changes in global climate characteristics, their grid system is too large to assess the impacts on major hydrologic parameters such as soil moisture and runoff on regional scales (Gleick 1987a). For such assessments, hydrological modelling with hypothetical climatic inputs, has been used. This latter approach has been criticized for some arbitrariness of assumptions of detailed climatic changes (Klemes 1983).

IPCC (1990) as well as others (McCabe and Ayers 1989; Nemec 1989) recommended to construct hypothetical scenarios to study characteristics of runoff responses to climate changes for particular areas. Such an approach has been adopted in this study dealing with simulated changes in the hydrologic cycle of a small basin, located at an intermediate northern latitude (47.5° N), arising from hypothetical changes in climatic inputs. Using a calibrated continuous simulation model, generally accepted ranges of variations in precipitation and air temperature inputs were adopted to assess the sensitivity of simulated streamflow in terms of annual, monthly and monthly-peak flows.

Study Area

Earlier studies of hydrologic impacts of climatic changes focussed on intermediate-to-large basins with intermediate annual mean temperatures (Nemec and Schaake 1982; Gleick 1987b; Bultot *et al.* 1988; McCabe and Ayers 1989). This study addressed the impacts of climatic input changes on the hydrology of a small basin with significant snowfall/snowmelt, low annual mean temperatures (4.8° C) and a high conversion of precipitation into streamflow. Under such circumstances, significant

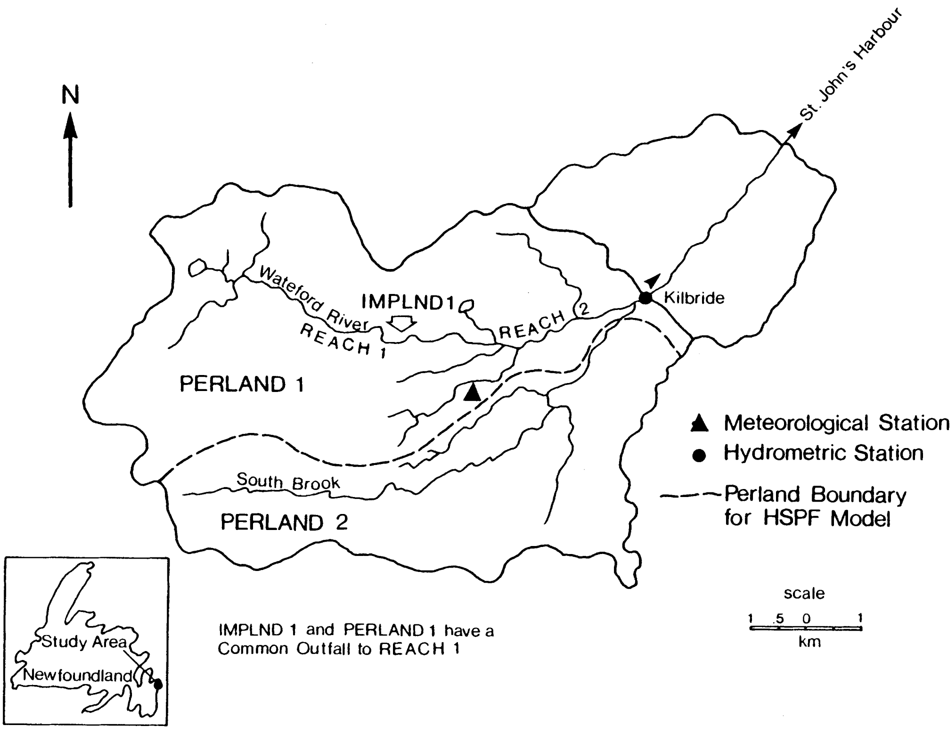

Fig. 1. Study area – Location map.

flood peaks may occur during snowmelt periods, particularly in coastal regions with increased probabilities of rainfall/snowmelt events over saturated or frozen grounds. Occurrences of such events may further increase as a result of climatic changes and concomitant greater variations in event characteristics, including an increased probability of warm spells and the resulting fast snowmelt (IPCC 1990). A description of characteristics of the Waterford River Basin follows.

Physiography and Land Use

The Waterford River Basin is located near St. John's, Newfoundland as shown in Fig. 1. The study area represents an upper part of the watershed upstream of the hydrometric station and measures about 53 km^2. Bedrocks in the study area are of the latest Precambrian age and are overlain by clastic sediments of slate-siltstone, sandstone, conglomerate and granitic rocks. About 90 % of the field area is covered by veneer of till. The soils have developed from coarse textured siltstone, slate, sandstone and acid volcanic rocks. Such soils can be characterized by rapid surface drainage, but poor-to-very poor internal drainage. The bedrock geology and soil characteristics indicate low infiltration and fast hydrological response (Ng and Marsalek 1989).

Land use in the basin (Ng and Marsalek 1989) comprises forests - 33 %, unproductive land - 29 %, urban land - 19 %, agricultural land - 11 %, and land in transition - 8 %. Land use features of the study area include forests, agricultural fields, urban developments, ponds, bogs, stream channels, and gravel pits. The imperviousness of the area was estimated at 5 % (2.65 km^2).

Climate

The climate in the study area is moderated by the sea. The annual precipitation of 1,514 mm is distributed, on the average, over 207 days in a year. Snowfall accounts for one third of the total precipitation. Evapotranspiration in the basin is rather low, between 250 and 350 mm annually (Yoxall 1980; Ng and Marsalek 1989). The average daily minimum and maximum temperatures are 1.0° C and 8.6° C, respectively, and the annual mean temperature is 4.8° C. The yearly solar radiation is 1,540 W m^{-2} and the yearly wind speed, all directions combined, is 24 km h^{-1}.

Streamflow

High monthly streamflows usually occur in the spring as a result of the combined effect of rainfall and snowmelt. The extreme minimum and maximum flows observed in the Waterford River at Kilbride from 1974 to 1984 were 0.15 m^3s^{-1} and 66.1 m^3s^{-1}, respectively. The long-term (1975-1985) mean annual flow was 1.65 m^3s^{-1}. Low flows usually occur from June to August. The hydrologic response of the study area is characterized by high conversion of precipitation into runoff (> 80 %) and a high ratio between high and low flows.

Modelling Methodology

Model Selection

Various hydrologic models have been used to study the impacts of climate change scenarios, depending on the study purpose and model availability. For assessing water resources management on a regional scale, water-balance models were found useful for identifying hydrologic consequences of changes in temperature, precipitation, and other climatic variables (Gleick 1986; Sanderson and Smith 1990). For the assessment of climate variability and change impacts on groundwater in the Arhus region of Denmark, Thomsen (1990) proposed to use the European Hydrological System (SHE) model.

In detailed assessments of surface flows, streamflow simulation models are used. Nemec and Schaake (1982) emphasized the suitability of deterministic models using daily or even shorter time steps. In their study of three river basins, they used the Sacramento Soil Moisture Accounting Model, because of its availability and the general acceptance in various climatic conditions. Similar selection criteria were applied in this study and led to the adoption of the Hydrologic Simulation Program-FORTRAN (HSPF) (U.S. EPA 1984), which was calibrated earlier for the study area (Ng and Marsalek 1989). Furthermore, the HSPF model uses first-order meteorological data inputs which can be easily altered to approximate climate change scenarios. This model was also recommended by Lumb and Linsley (1971) for studies of rainfall augmentation impacts on streamflow, which is a phenomenon similar to increased precipitation due to climate warming.

Model Description

The HSPF model is a comprehensive continuous simulation model for predicting the watershed hydrological response. It allows detailed simulation of streamflow, flow hydraulics, water quality processes, contaminant transport and lake reservoir dynamics. The model employs information on the time history of precipitation; wet bulb and dry bulb air temperatures; wind movement; solar radiation; actual evaporation; cloud cover; and, land use characteristics including patterns, soil characteristics, and vegetation. It simulates hydrological and chemical processes that occur in the watershed and in-stream transport. The result of such simulation is a time history of the quantity and quality of runoff.

A detailed description of the HSPF model is available in the literature (U.S. EPA 1984). A brief overview of the HSPF hydrological components, with emphasis on evapotranspiration and snowmelt, is given below to facilitate later discussion of modelling results.

Depending on the air temperature, two forms of precipitation are considered in the HSPF model, rain or snow. Once snow begins to accumulate on the ground, the snowpack accumulation and melt calculations consider the following sources of heat: net radiation heat (both longwave and shortwave), sensible heat from the air, latent heat transfer by condensation of moist air on the snowpack, sensible heat from falling rain, and heat from the underlying ground. Energy calculations are used to calculate the temperature of the snowpack and for temperatures above 0° C, melting takes place. Net heat losses cool the snowpack and produce negative heat storage. Incoming heat from the ground melts the snowpack from the bottom. As a modelling option, ice in or frozen ground under the snowpack can be simulated.

Precipitation is subject to interception representing moisture storage (abstraction) above the overland flow plane. Outflow from interception storage enters the surface detention storage of land segments and is added to the existing storage to make up water available for infiltration and runoff.

Depending on physiographic and hydrologic conditions, water from the detention storage may stay on the overland flow plane and run off or infiltrate into the ground. Water infiltrating into the ground may enter the following storages - the upper soil moisture zone, the lower soil moisture zone, the active groundwater storage and the deep (inactive) groundwater storage. The first three storages may contribute to interflow or baseflow, but deep percolation to the inactive groundwater storage is considered lost from the simulated system.

Evapotranspiration extracts water from various storages, including interception, upper and lower zone storages, active groundwater and baseflow, and returns it to the atmosphere. To determine evapotranspiration (*ET*), first the potential *ET* must be established from the input data series. Secondly, the actual *ET* is calculated by trying to meet the demand from five sources in the order described below.

The first source is the active groundwater outflow or baseflow subject to *ET* by riparian vegetation. The user can specify the fraction of *ET* that can be withdrawn from the baseflow. The remaining potential *ET* exerts its demand on the water in the interception storage until it is exhausted. Next, the *ET* is drawn from the upper zone in proportion to the ratio of moisture stored to the nominal moisture capacity. The actual *ET* from the active groundwater is simulated by specifying the fraction of the remaining potential *ET* that should be drawn from this source. The lower soil zone is the last storage from which *ET* is drawn. This *ET* component depends on vegetation transpiration, governed by the vegetation type, the depth of roots, density of vegetation cover, and the stage of plant growth along with moisture characteristics of the soil.

This model is generally applied using a specific time interval related to the process time scale. Once the model has been calibrated, process parameters are easily modified to produce outputs corresponding to various scenarios. The input data time series can be altered in the model before they are used in simulation. Such flexibility in data handling is ideal for studies of climatic scenarios in a given watershed. Thus, the HSPF model is well suited to address the impacts of climate warming on streamflow from the area studied.

The HSPF model was applied to the Waterford River Basin earlier (Ng and Marsalek 1989) and, in this study, the earlier model set-up was adopted without major changes. Basic characteristics of this model set-up appear in Table 1 and Fig. 1.

Table 1 – HSPF Model Set-up – Land Segment Characteristics

Characteristic	Pervious Area		Impervious Area
Segments	PERLAND 1	PERLAND 2	IMPLND 1
Latitude (degrees)	47.5° N	47.5°N	47.5°N
Area (km^2)	30.25	20.13	2.65
Elevation (m ASL)	150	110	110
Overland Flow Length (m)	1,750	1,000	1,000
Overland Flow Mean Slope	0.015	0.015	0.010
Forest Cover (%)	34.0	34.0	0.0

Model Calibration

For the study area, the HSPF model has been calibrated using 29 months of historical hourly data. This calibration started with the annual water balance, followed by seasonal and monthly flows, and event peak flows. A brief summary of calibration results follows; detailed results can be found elsewhere (Ng and Marsalek 1989). The total streamflow over the calibration period was reproduced by the calibrated model within 1 per cent of the observed streamflow. Monthly streamflows were reproduced, on the average, within ± 15 per cent of the observed ones. The goodness of fit was adversely affected by unfavourable results in several months with low streamflow. Finally, event flow peaks, greater than 10 m^3/s, were reproduced on the average within five per cent. Following this calibration, the HSPF model was considered to be appropriate for simulation experiments with various climatic data.

Approximation of Climatic Changes

Hydrometeorological input data required by the HSPF model for streamflow simulation, as prepared in the earlier study (Ng and Marsalek 1989), included hourly air temperatures (wet and dry bulb), wind speed, precipitation, solar radiation, and daily evaporation (lake and pan) data. Such historical data were then used to produce hypothetical input data approximating perceived climatic changes.

Accurate forecasts of climatic changes resulting from global warming are beyond the current understanding of these processes. Consequently, it is necessary to rely on crude estimates of ranges of expected changes in climatologic data. Such ranges are typically expressed by monthly or annual means of temperature and precipitation (Sanderson and Smith 1990). For estimation of impacts of climatic changes on water resources, the annual data are inadequate and further discretization of such data into seasonal, monthly, daily and even shorter intervals may be required. With each finer discretization step, the inherent uncertainties are increased, but there are no better alternatives available at this time.

In terms of ranges of annual climatic changes, it is generally agreed that temperatures may increase from 1° to 4° C, and, depending on the latitude, and precipitation may vary by as much as ±25 % (Nemec and Schaake 1982; Gleick 1987a; WMO 1987; McCabe and Ayers 1989; IPCC 1990). These annual changes were distributed during the year by various methods. For example, Sanderson and Smith (1990) used a GISS (Goddard Institute of Space Studies) scenarios of predicted monthly changes in temperature and precipitation. The resulting monthly variations in the temperature and precipitation can be described by relatively low coefficients of variation of 0.18 and 0.08, respectively. Another global climate model, used by Manabe and Stouffer (1980), indicated somewhat greater variations of temperature increases during the year, with the highest increases found during the winter months.

Nemec and Schaake (1982) assumed constant distributions of climatic changes and multiplied historical precipitation records by constant factors and adjusted historical temperatures by constant increments. The latter approach was adopted in this study by applying constant factors or increments to historical hourly data. In simulation scenarios, the hourly temperatures in the available data base were increased by 1, 2, 3 and 4° C, respectively, and precipitation was varied by plus/minus 5, 10 and 20 %, respectively. The above changes should cover the currently anticipated range of climatic changes.

In all scenarios, it was assumed that dew point temperatures would increase by the same amount as the air temperatures. The amount of increase or decrease in precipitation would occur either as rainfall or snow, depending on air temperatures. Uncertainties in predicting changes in other meteorological parameters such as wind speed, solar radiation, cloud cover, and actual evaporation are even greater and cannot be evaluated, on regional scales, by the current atmospheric models. For this reason, all the previously mentioned parameters were assumed to be the same as under the existing conditions and effects of their changes were assumed negligible in the study area.

Results and Discussion

Simulation results for hypothetical climate change scenarios were evaluated with various temporal detail, starting with the water year, followed by seasonal or monthly periods, and by the largest monthly events. The presentation of results follows the same order. It should be also emphasized that any observations made in this section are based on a relatively short period of simulations and would require further substantiation in the future.

The simulated water-year streamflows for various changes in precipitation and temperature inputs are shown in Fig. 2. Several earlier modelling studies indicated that warming led to higher evapotranspiration, reduced soil moisture and reduced runoff/streamflow (Gleick 1987b; Bultot *et al.* 1988; McCabe and Ayers 1989; Sanderson and Smith 1990). The results produced in this study for the maximum temperature increase of 4° show similar impacts on the actual evapotranspiration which increased by about 6 %. However, since *ET* represents just about one sixth of annual precipitation in the study area (Yoxall 1980), the annual streamflow was reduced by about 1 % and such a change was considered insignificant. These results were accepted as plausible because of low evapotranspiration in the study area and seemed to be further supported by Karl and Riebsame (1989) analysis of historical data showing minimal effects of recent temperature fluctuations on streamflow. Furthermore, indirect effects of increased CO_2 concentrations are manifested by increased plant stomatal resistance and concomitant reductions in evapotranspiration (Aston 1984). Thus, for the basin studied, the impacts of higher temperatures on annual streamflow appear to be negligible.

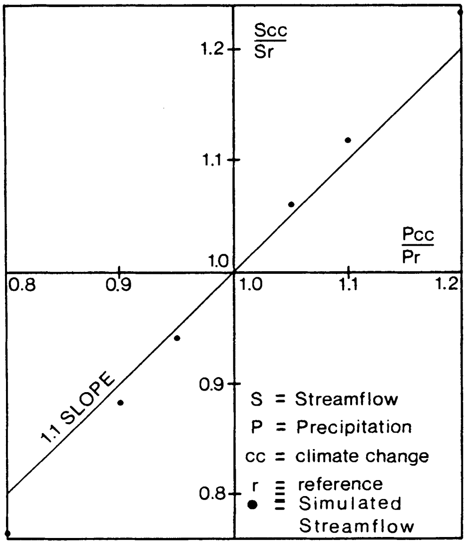

Fig. 2. Annual streamflow changes simulated for various changes in annual precipitation.

Streamflow simulated for the study area varied linearly with changes in precipitation. This linearity in the study basin response was explained by high conversion of precipitation into streamflow and negligible changes in evapotranspiration with varying precipitation. This finding is well supported by the data reported by Lumb and Linsley (1971) showing that, in basins with average annual runoff greater than 60 % of average annual precipitation, the added precipitation is converted into runoff and there are no changes in evapotranspiration. Thus in terms of annual streamflow, the simulated changes are comparable to the changes in precipitation which from the water management point of view are not very significant in this particular case.

Variations of simulated monthly and seasonal streamflow resulting from changes in temperature and precipitation inputs are shown in Fig. 3. Increased temperatures affected the distribution of annual streamflow during the period studied. The earlier studies, for larger basins, reported earlier and faster snowmelt and increased runoff during the winter months (Gleick 1987b; McCabe and Ayers, 1989; Bultot *et al.* 1988). This tendency was noted for the study area for the winter of 1981/82, when the winter runoff hydrograph occurred about 2.5 month earlier for the temperature increase of 4° C scenario and about one month earlier for the + 1° C scenario. The resulting April 1982 streamflow represented the highest value during the relatively short study period and equalled about twice the reference run monthly streamflow.

In the winters of 80/81 and 82/83, no advancement of winter runoff was noticed, because there was little precipitation stored in the catchment snowpack and the

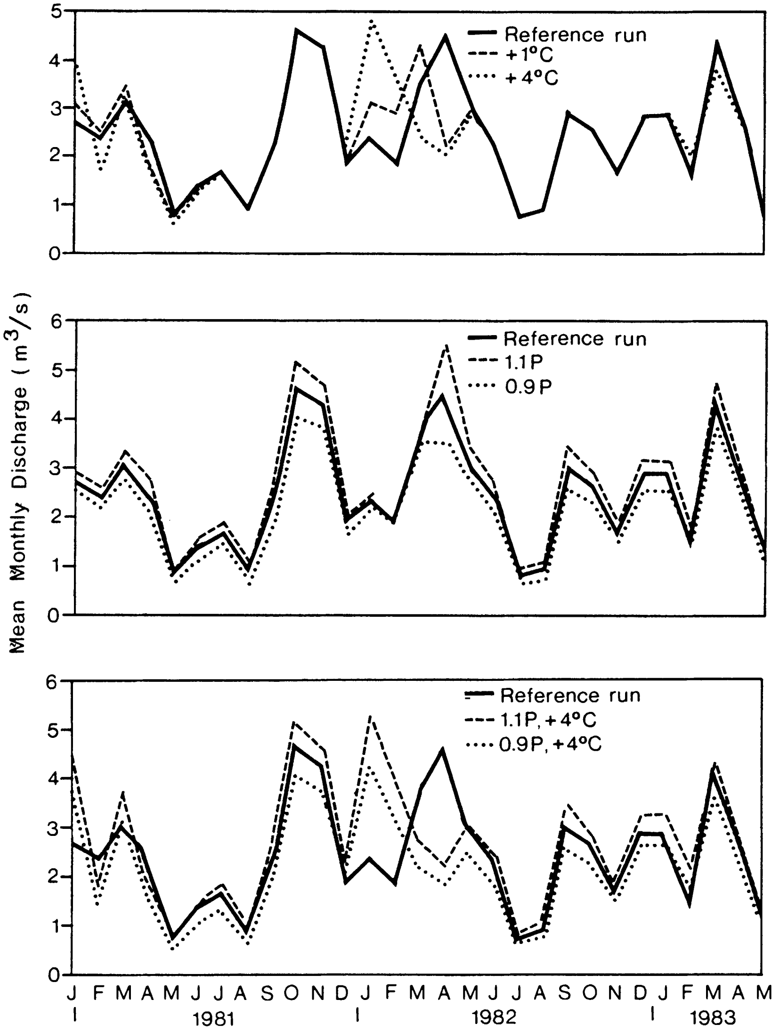

Fig. 3. Monthly streamflow simulated for various climatic scenarios.

incoming precipitation was running off quickly either as snowmelt or runoff. Only in January of 1982 there was a large snowfall that under the reference run conditions, was stored on the catchment, but started to melt slowly with a temperature increase of 1° C and quickly for an increase of 4° C. In the remaining months, no significant impacts of temperature on monthly streamflows were noticed. Thus in the study area, warming may cause earlier winter runoff, but only when there is precipitation stored in the catchment snowpack.

Variations in precipitation (no change in temperature) resulted in linearly proportional changes in streamflow. This proportionality results from high conversion of precipitation into streamflow and the described basin response is in agreement with the results of Lumb and Linsley (1971) and Bultot *et al.* (1988). Thus, increases or decreases in precipitation from 0 to 20 % were reflected by comparable changes in streamflow. In winter months, some redistribution of increased precipitation may occur. Additional precipitation occurs in the form of snow which is stored in the catchment and subsequently melted. This process may lead to streamflow increases significantly exceeding (up to a factor of two) the uniform increase in precipitation in the month with the highest snowmelt. Such a streamflow magnification can be seen in Fig. 3 for April, 1982.

Finally, it was of interest to examine the impact of combined changes in precipitation and temperature on monthly streamflow. For this purpose, streamflow simulations were done for 24 combinations of changes in temperature (+1, 2, 3, and 4° C) and precipitation (±5, ±10, and ±20 %). For clarity, only two important change scenarios are shown in Fig. 3. These scenarios, A and B, assume a temperature increase of 4° C and precipitations of 0.9 P_{ref} and 1.1 P_{ref}, respectively. Both scenarios led to some redistributions of annual streamflow, particularly during the winter of 1981/82. Scenario A (+4° C, 0.9P) led to generally lower monthly streamflows, with the exception of January of 1982 when early snowmelt produced streamflow almost twice as large as that simulated in the reference run. Scenario B also produced higher streamflows except in March and April, 1982, when early snowmelt removed stored precipitation and the reference run produced higher streamflows resulting from delayed snowmelt.

The preceding discussion indicates that the climatic input changes studied may lead to a significant redistribution of the annual streamflow, with snowmelt playing an important role. Under specific conditions, a fast melt of the catchment snowpack can be induced and this would lead to increased winter/spring monthly streamflows that will further affect event peak flows occurring during these months, as discussed below. This appears to be the only significant seasonal change observed as a result of climatic input changes – earlier and faster snowmelt, followed by reduced streamflows in the later spring months (March-May). Such results were affected by the climatic conditions in the study area corresponding to a relative cool and temperate climate moderated by the sea and the temperature records specific to the study period. Winter temperatures often drop very little below the freezing point and the simulated increases in temperature may lead to above-zero temperatures and concomitant snowmelt.

Annual and seasonal streamflows are of great interest for such water management issues as basin yield and water use, but for other issues, such as flood protection, actual river discharges are needed. Consequently, the analysis of simulation experiments was extended to individual events and their peak flows were evaluated. It is recognized that no reliable physical basis exists for extending the perceived climatic changes to short time intervals. Furthermore, the changes will not be uniformly distributed, but may involve significant variations for individual

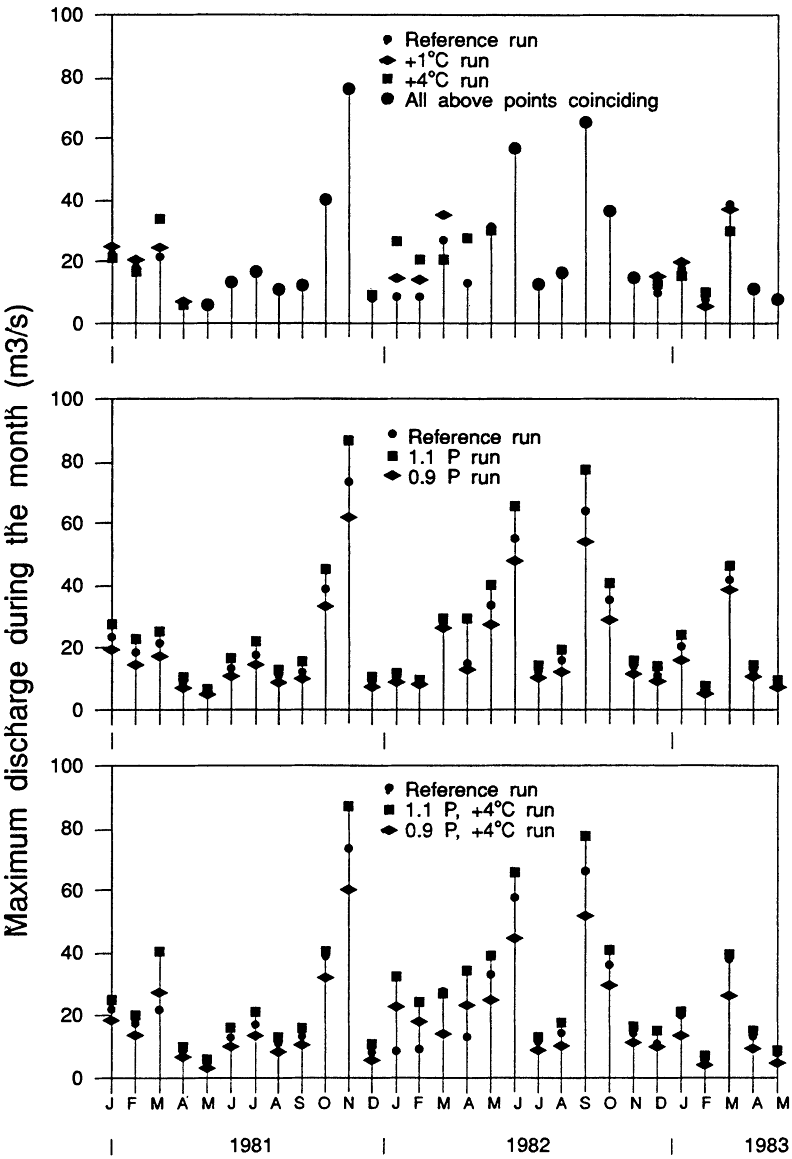

Fig. 4. Monthly flow peaks simulated for various climatic scenarios.

events. With these precautionary statements, the results of simulation experiments were evaluated for maximum monthly peak flows. In the interpretation of such data, only the trends are emphasized, recognizing that magnitudes of peak flow changes are largely simulation artifacts depending on many hypothetical assumptions.

Changes in monthly maximum flows, resulting from variations in temperature and precipitation, are shown in Fig. 4. Effects of temperature increases are detected only for winter months. In general, higher temperatures produced flows that were significantly higher (up to three times) than the flows simulated in the reference run. These changes were found only for relatively low flows which would not be important in flood analysis. In the study basin, the highest ranked flows all occurred outside of the winter season and their simulations were not affected by increased temperatures. In basins where the annual peaks may be generated during the winter months, warming could produce increases in flood peaks. Such increases in flow peaks would not be caused just by snowmelt, but as reported by Bengtsson (1985), by the combined occurrence of rainfall/snowmelt over saturated or frozen ground.

Enhanced precipitation resulted in increased flow peaks. Such changes were proportional to the changes in precipitation, with the coefficient of proportionality equal to two. Thus, for precipitation increased by 10 %, the flow peaks would increase by about 20 %. The combined effects of temperature increase and precipitation variations were also examined (Fig. 4). These effects were primarily controlled by the changes in precipitation, with the temperature playing a minor role during the winter months.

In the overall evaluation, the monthly maximum flows were found to be relatively insensitive to changes in temperature changes. Their sensitivity to changes in precipitation was characterized by an amplification factor of two.

Summary and Conclusions

The sensitivity of streamflow simulations to changes in climatic input data was studied for a small catchment with significant snowfall and a high conversion of precipitation into streamflow (> 80 %). In simulation experiments, both temperature and precipitation data were uniformly varied over the simulation period of 29 months and the effects of such variations on simulation results were evaluated in terms of annual streamflow, seasonal/monthly streamflows, and monthly maximum peak flows.

The annual streamflow was barely affected by changes in temperature – reductions of up to several per cent resulted from increased evapotranspiration. Variations in precipitation resulted in linearly proportional changes in annual streamflow, without any amplification. Seasonal streamflow was sensitive to both temperature and precipitation changes. Higher temperatures resulted in earlier and faster snowmelt, during a winter with a significant water storage in a snowpack, or in higher proportions of precipitation in the form of rainfall and faster runoff from the basin. Monthly maximum peak flows were not excessively sensitive to climatic input changes. Increases in precipitation led to somewhat amplified (two times) increases in flow peaks. While higher temperatures led to higher peaks during the winter months for the basin studied, such increased peaks were still smaller than the maximum peaks observed in this basin outside of the winter period.

References

Aston, A. R. (1984) The effect of doubling atmospheric CO_2 on streamflow: A simulation, *Journal of Hydrology, Vol. 67,* pp. 273-280.

Bengtsson, L. (1985) Characteristics of snowmelt induced peak flows in a small nothern basin, *Nordic Hydrology, Vol. 16,* pp. 137-156.

Bultot, F., Coppens, A., Dupriez, G. L., Gellens, D., and Meulenberghs, F. (1988) Repercussions of a CO_2 doubling on the water cycle and on the water balance – a case study for Belgium, *Journal of Hydrology, Vol. 99,* pp. 319-347.

Cohen, S. J. (1986) Impact of CO_2 induced climate change on water resources in the Great Lakes Basin, *Climatic Change, Vol. 8,* pp. 135-153.

Gleick, P. H. (1986) Methods for evaluating the regional hydrologic impacts of global climatic changes, *Journal of Hydrology, Vol. 88,* pp. 97-116.

Gleick, P. H. (1987a) Regional hydrologic consequences of increases in atmospheric CO_2 and other trace gases, *Climatic Change, Vol. 10* pp. 137-161.

Gleick, P. H. (1987b) The development and testing of a water balance model for climate impact assessment: modeling the Sacramento Basin, *Water Resources Research, Vol. 23(6),* pp. 1049-1061.

Intergovernmental Panel on Climate Change (IPCC) (1990) *Climate Change.* The IPCC Scientific Assessment (Edited by J. T. Houghton, G. J. Jenkins, and J. J. Ephraums), Cambridge University Press, Cambridge, U.K.

Karl, T. R., and Riebsame, W. E. (1989) The impact of decadal fluctuations in mean precipitation and temperature on runoff: A sensitivity study over the United States, *Climatic Change, Vol. 15,* pp. 423-447.

Kite, G. (1989) Use of time series analysis to detect climatic change, *Journal of Hydrology, Vol. 111,* pp. 259-279.

Klemes, V. (1983) Climatic change and the planning of water resource systems, Proceedings of the 6th Canadian Hydrotechnical Conference, June 2-3, Canadian Society for Civil Engineering, Ottawa, Ontario, Canada, pp. 485-500.

Klemes, V. (1990) Sensitivity of water resource systems to climatic variability, Proceedings of the 43rd Annual Conf. of the Canadian Water Resources Association, May 16-18, Penticton, British Columbia, Canada, pp. 233-241.

Lumb, A. M., and Linsley, R. K. (1971) Hydrologic consequences of rainfall augmentation, *Journal of the Hydraulics Division, Proceedings of the American Society of Civil Engineers, Vol. 97(HY7),* pp. 1065-1080.

Manabe, S., and Stouffer, R. J. (1980) Sensitivity of a global climate model to an increase of CO_2 concentration in the atmosphere, *Journal of Geophysical Research, Vol. 85(C10),* pp. 5529-5554.

McCabe, G. J., Jr., and Ayers, M. (1989) Hydrologic effects of climate change in the Delaware River Basin, *Water Resources Bulletin, Vol. 25(6),* pp. 1231-1242.

Nemec, J. (1989) Implications of changing atmosphere on water resources, Proceedings of the WMO Conference, June 27-30, Toronto, Canada, WMO-710, pp. 211-223.

Nemec, J., and Schaake, J. (1982) Sensitivity of Water Resources System to Climate Variation, *Hydrological Sciences, Vol. 27(3),* pp. 327-343.

Ng, H. Y. F., and Marsalek, J. (1989) Simulation of the effects of urbanization on basin streamflow, Water Resources Bulletin, No. 25(1), pp. 117-124.

Sanderson, M., and Smith, J. (1990) Climate change and water in the Grand River, Ontario, Proc. of the 43rd Annual Conf. of the Canadian Water Resources Association, May 16-18, Penticton, British Columbia, pp. 243-261.

Thomsen, R. (1990) Effect of climate variability and change in groundwater in Europe, *Nordic Hydrology, Vol. 21,* pp. 185-194.

U.S. Department of Energy (1980) Carbon dioxide effects research and assessment Program, Part 1: The global carbon cycle and climate effects of increasing carbon dioxide, DOE/EV-0094, UC-11, 99 pp.

U.S. Environmental Protection Agency (1984) Users manual for Hydrological Simulation Program – Fortran (HSPF) EPA-600/3-84-066, Environmental Research Laboratory, Athens, Georgia, 767 pp.

WMO (1987) Water resources and climate change – sensitivity of water-resources systems to climate change and variability, World Climate Programme applications, WMO-No. 247, Geneva, Switzerland, 51 pp.

Yoxall, W. H. (1980) The Water Balance in Newfoundland, Water Resources Bulletin, Vol. 16, pp. 1030-1033.

First received: 9 August 1991
Revised version received: 20 May, 1992
Accepted: 1 June, 1992

Address:
National Water Research Institute,
867 Lakeshore Road,
Burlington, Ontario,
Canada L7R 4A6.