

Climate impacts on the ice regime of an Atlantic river

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Abstract Although the sensitivity of river ice processes to climatic inputs is well known, there is very little information on how a changing climate can affect the severity and frequency of ice jam events and their numerous ecological and socio-economic impacts. The present study adds to this information by examining the ice regime of the Southwest Miramichi River, New Brunswick, and identifying recent trends that may be linked to concomitant climatic variations. The timing of freeze-up and breakup, as well as the thickness of the winter ice cover, do not exhibit significant temporal trends. However, spring ice jamming is becoming more severe, and there is increasing potential for devastating mid-winter breakup events. These findings are consistent with increasing rainfall and snowfall amounts, as well as increasing river flows, during the winter and early spring. Unlike in most parts of Canada, slight cooling during the winter months was detected, consistent with cooling trends found elsewhere in Atlantic Canada. Implications for adaptation and infrastructure design are discussed.

Keywords Climate; flow; ice jamming; ice regime; river; trend

Introduction

Ice is present in nearly every Canadian river, for a period that ranges from days to many months. Whether moving or stationary, it interacts with the river flow in various ways, resulting in multiple impacts on the economy and ecosystem, and often posing a major threat to riverside communities. Extreme events resulting from breakup ice jamming are responsible for a large proportion of such impacts. They include flooding, damage to property and infrastructure, interference with navigation, and inhibition of hydropower generation. In New Brunswick, where detailed damage records are available, it has been found that ice jams cause one-third of all flood events but are far more destructive than open-water floods because they account for two-thirds of all flood damage (Humes and Dublin 1988).

Equally important are the many ecological impacts of river ice, which arise from the intimate relationship between ice processes and riverine ecosystems (Prowse 2000). Extreme ice-jam events are again major causes of ecological impacts that can be both beneficial and detrimental. For instance, ice-jam flooding provides essential replenishment to the multitude of lakes and ponds characteristic of the northern Canadian deltas, which are havens for wildlife, especially for waterfowl and aquatic animals. (Peace–Athabasca Delta Project 1973; Marsh and Hey 1989). On the other hand, flooding caused by ice jams and the surges produced by their release can result in severe fish mortality and loss of spawning grounds. Because ice is almost ubiquitous on Canadian rivers, aquatic species have adapted their life cycles so as to take advantage of, or cope with hazards posed by, various phases of the annual ice regime.

In response to the increasing recognition of the economic and ecological significance of river ice, there has also been a rising need to forecast how it will react to changes resulting from global warming (e.g. Fitzharris *et al.* 1995). Climate, particularly temperature and precipitation, shapes the flow hydrograph and the stream morphology, which in turn govern

river ice processes. Such processes, and especially breakup, which involves the decay, fracture, transport, jamming, and eventual clearance of the river ice cover, are very sensitive to climatic inputs. Climate change can result in greatly modified ice regimes, with the potential for a variety of ecological impacts. Therefore, local hydroclimatic conditions need to be considered when developing environmental adaptation strategies.

The same is true for planning and designing infrastructure in the vast majority of Canadian rivers. Because the design life of infrastructure often extends to 50 years and beyond, the effect of climate change on the local ice regime must be addressed. This is prudent not only with respect to the safety of the infrastructure itself, but also as a safeguard against amplified environmental impacts in the future. The present study was motivated in part by climatic concerns that arose during an ongoing investigation into the interaction of ice with bridge structures (Beltaos *et al.* 2001; Sullivan and Beltaos 2000). In addition to analytical and modeling work, this investigation includes detailed observations at two bridge sites on the Southwest Miramichi River, respectively located upstream and downstream of the Blackville hydrometric gauging station. The latter is the main source of flow discharge information for the entire reach. Archived hydrometric data have been analyzed to quantify the ice regime, using the methodology developed by the author (see Beltaos *et al.* 1990), and to examine trends in relatively recent hydroclimatic conditions.

The results of the study are reported in two parts. The first part deals with ice processes, as revealed by the analysis of the hydrometric station records, and includes comparisons with previous findings and current theoretical concepts. The processes of ice cover formation, ice thickness growth and decay, onset of breakup, ice-jam severity and frequency are described and quantified. The second part delves into the historical aspect of the study, by examining trends in ice processes over the period of record and investigating their relationships to concomitant trends in hydrologic and meteorological variables. The main findings are then summarized and implications for future adaptation are discussed.

Background information

To date, the links between climate and river ice remain largely unexplored. Relevant studies have focused on simple indicators, such as changes in freeze-up and breakup dates, duration of the ice season, and ice cover thickness. Evaluation of more complex occurrences, such as the magnitude and frequency of ice-jam floods, depends on numerous factors driven by climate but interacting through a variety of geophysical processes such as ice mechanics, micro-meteorology, hydrology and hydraulics. Only two studies of this kind can be cited at present, respectively pertaining to Eastern and Western Canada (Beltaos 2002; Beltaos *et al.* 2002). The former study suggests an increasing frequency and severity of ice jams for the upper Saint John River (NB), while the latter points to the opposite trend for the lower Peace River in the Peace–Athabasca Delta (Alberta).

The province of New Brunswick is not only prone to river ice impacts, but also exhibits strong climatic variability, being situated between the Canadian mainland on the north and west, and the Bay of Fundy and the North Atlantic on the south and east. Abraham *et al.* (1997) contains preliminary analyses of runoff and river ice data that reveal evidence of a mild trend towards fewer days with ice in northern New Brunswick rivers. However, there appeared to be no change in the number of days that ice is present in southern New Brunswick rivers. The results exhibited the strong geographic variability of climate in Atlantic Canada, and might indicate that the impacts of climatic change on water resources would not be uniform across Atlantic Canada, or even within New Brunswick. Brimley and Freeman (1997) suggested a linkage between winter ocean currents and the formation of ice on the streams of Atlantic Canada. They stated that the moderating effect of the ocean that

commonly kept streams in coastal regions (especially in Nova Scotia and Newfoundland) open longer, seems to have diminished since the 1960s.

Hare *et al.* (1997a) determined that mean annual precipitation and streamflow in the Saint John River Basin appeared to have become more variable since 1950, but no overall trends were observed. They also noted that freshets have generally arrived earlier since 1972. Only a small rise of spring temperatures was detectable, but snowy or wet winters, coupled with greater variability, have caused earlier thaws and several major open-water and ice-jam flooding events. Hare *et al.* (1997b) stated that one-day heavy rain or snowfall events have increased over the Saint John River Basin. They warned that the potential for severe storms to occur simultaneously with the spring thaw and the onset of the freshet confirms a need for precautionary measures to lessen the likelihood of future flood damages.

To examine the causes of three recent winter breakup events that occurred on the upper Saint John River, located in NW New Brunswick, Beltaos (2002) carried out a hydroclimatic analysis of local archived data. Using long-term climatic and flow records, he found that a small rise in winter air temperatures over the past 80 years had resulted in a large increase in the number of mild winter days. As a result, the amount of winter rainfall (as opposed to snowfall) had increased, augmenting flows sufficiently to cause breakup of the ice cover. In addition, flow peaks were found to have increased during April, further increasing the risk of spring ice-jam flooding.

Little is known, however, about hydroclimatic aspects of the ice regime of other rivers in New Brunswick, especially streams that are more typical of those whose drainage basin lies in the central portion of the province.

Description of study reach and local ice processes

The Southwest Miramichi River originates in the highlands of west-central New Brunswick and flows in a generally southeasterly direction before turning to flow approximately 105 km in a more northeasterly direction through the New Brunswick Lowlands to its confluence with the Northwest Miramichi River. The Northwest and Southwest Miramichi Rivers join to form the 26-km-long tidal Miramichi River that flows northeast to empty into the Miramichi Bay (Beltaos *et al.* 2001).

The study reach (Figure 1) is a part of the lower portion of the Miramichi River basin, whose total area of 11,700 km² renders it the largest river basin entirely within the province of New Brunswick. A significant portion of the province's population resides within the basin, mostly in narrow strips along the lower Miramichi and Southwest Miramichi rivers. Most of the basin is forested. Several tributary streams flow eastwards from the New Brunswick Highlands, a rugged area of metamorphic rocks highly resistant to erosion. The streams combine to form major rivers in the eastern portion of the basin, which is within the New Brunswick Lowlands, an area of low relief and flat or gently dipping sedimentary rocks (Beltaos *et al.* 2001).

The gauge site (Figure 1) is located approximately 39 river kilometers upstream of the confluence with the Northwest Miramichi River, and some 65 km from Miramichi Bay. In the vicinity of the gauge site, the river consists of a single, mildly curved channel 100–150 m wide, depending on water surface elevation and proximity to the gauge (Figure 2). Water surface elevations along the reach defined by the cross-sections indicated in Figure 2 were surveyed on 13 July 1999, and the average slope is calculated as 0.36 m/km. Also, using bathymetric data from the same two sections and gauge rating tables, the reach hydraulics can be determined for open-water conditions. As is often the case (e.g. see Beltaos 1995), the bed Manning roughness coefficient decreases from a high value of 0.075 at very low flow (50 m³/s) to the more familiar value of 0.030 at a very high flow of 2000 m³/s.

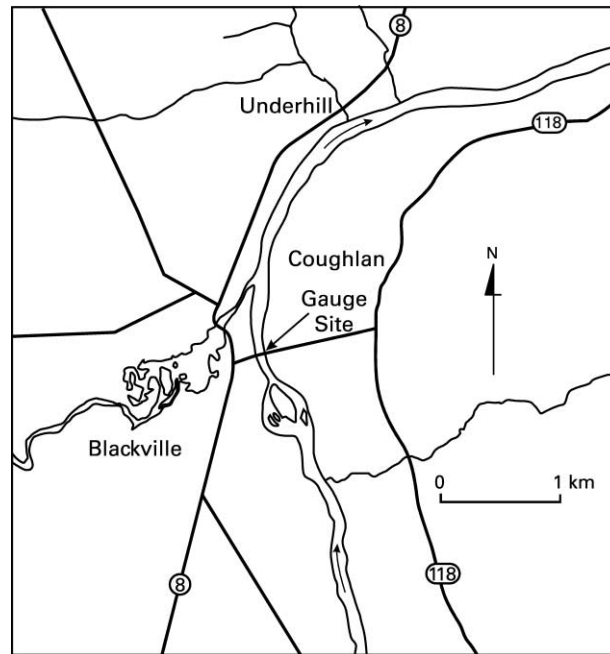


Figure 1 Gauge reach

Typically, an ice cover forms at Blackville in December and breaks up in April. Freeze-up observations in 1998 (Beltaos *et al.* 2001) revealed that the ice cover progressed rapidly in the upstream direction from the flat, tranquil reach of the river above its confluence with the Northwest Miramichi. Here, the cover had the smooth appearance of sheet ice that typically forms under conditions of surface juxtaposition of ice floes. The upstream advance of the ice edge slowed in a reach of rapids between 6 and 9 km below the gauge site, due to shoving and thickening, manifested in a cover with a rough surface. Shortly after the ice cover edge progressed past the rapids reach, it was again stalled by bridging-over at the island above the gauge, which cut off the greater part of the ice supply. Eventually, however, the entire reach freezes over, and, by early January, the ice cover is 30 cm to 40 cm thick near the gauge site (Beltaos *et al.* 2001).

Following freeze-up, the stage drops with the decreasing flow, attaining a relatively steady value until mid- or late March, when it again begins to rise in response to snowmelt that is often accompanied by rainfall. Breakup is usually initiated in April and typically lasts

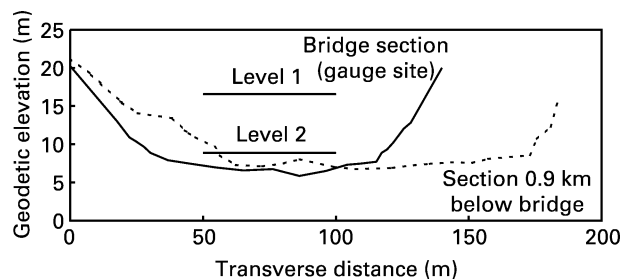


Figure 2 River cross-sections in the vicinity of the gauging station. Level 1 = highest ice-influenced stage on record (1970); level 2 = stage on date of survey, 12 July 1999; a typical winter stage is about 0.8 m higher than the level 2

for several days. In some years, the initial rise in flow and stage does not lead to breakup but is followed by a sustained “plateau” or relatively constant-stage phase, before there is additional runoff that leads to breakup. This feature has also been encountered on the Restigouche River (Beltaos and Burrell 1992), and probably results from persistent but modest runoff, largely due to snowmelt, which is not quite capable of dislodging the ice cover. If the final runoff event does not materialize quickly, the ice cover decays by thermal effects, sometimes to the point of near-disintegration. This situation produces a tame breakup event, known as “thermal” to distinguish from the more common “mechanical” breakup, where the mechanical properties of ice, despite some decay, still play a significant role in the process.

The study reach does experience occasional winter thaws. It is not unusual to encounter two or more such thaws before the final spring thaw that leads to ice breakup. Typically, the winter thaws last for a few days and attain maximum daily mean temperatures of a few degrees C. They happen more frequently in March, but they can also occur in January and February; as a rule, they do not produce a great deal of runoff. Consequently, winter breakup is rare, unlike in other parts of Atlantic Canada and in the New England states. When they do occur, however, winter breakups can be devastating (e.g. the major floods of January 1909 and of February 1970; Beltaos *et al.* 2001). Gauge records going back to 1962 indicate that December rainfall after the initial freeze-up can produce sufficient runoff to dislodge the newly formed ice cover (December 1964, 1969, 1983 and 1990), but these late-fall breakups are far more benign than the mid-winter ones, owing to relatively low river flow and thin ice cover.

Data sources

Historical daily flow data and chart-recordings of river stage at the Blackville gauge, operated by the Environmental Monitoring section of Environment Canada (formerly Water Survey of Canada), have been kindly provided to the author by the Fredericton office. After 1995, the stage record is available in digital format, at hourly intervals. Detailed flow and stage data are available after the 1961–1962 ice season. Daily-flow records go back to 1918 (e.g. see Environment Canada’s CD named “HYDAT”), although there are early periods of missing data. Stage and flow records are used to determine freeze-up/breakup times and related quantities (e.g. see Beltaos *et al.* 1990).

An important, although less well-known, gauge-related archive is the set of discharge-measurement notes. These are specially designed forms, used by field crews to record raw data when they carry out current-metering. During the winter, the distance from the water surface to the bottom of the ice cover is recorded at each borehole and subtracted from the total water depth to determine the flow depth. Typically, 20 or so holes are drilled for each discharge measurement, and the corresponding data can be very useful in estimating the average solid-ice thickness across the river. Complications arise when there are frazil deposits under the solid ice layer. Where the frazil deposit does not extend all across the channel width, it is often possible to identify the solid-ice portions because a frazil deposit tends to be much thicker than a solid-ice cover.

Ice processes

Ice thickness growth

The growth of the ice-cover during the winter is illustrated in Figure 3, where average thickness obtained during any one winter-flow measurement is plotted against the day of the year – for all the years for which data are available (1918–1939; 1962–1999). Figure 3 suggests that the thickness attains maximum season values of 0.5–0.8 m towards the middle

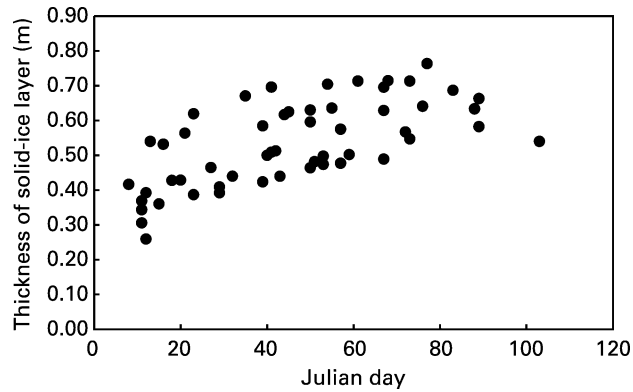


Figure 3 Ice thickness growth during the winter season

of March. Although Figure 3 provides a quick estimate of ice thickness as a function of time, improved prediction can be made in terms of the accumulated degree-days of frost (ADDF), as illustrated in Figure 4, where the solid line adheres to a Stefan-type equation:

$$h = 0.020\sqrt{(\text{ADDF})} \quad (1)$$

in which h = thickness of solid-ice cover in meters, and ADDF is expressed in °C-days. The numerical coefficient is somewhat higher than what is recommended for an “average river with snow” (0.014–0.017, e.g. see Beltaos, 1995).

Ice thickness decay

As winter gives way to spring, positive heat fluxes from the atmosphere above, and from the water beneath, the ice cover begin to develop and cause melting. However, sufficient data on heat fluxes are rarely available for a rigorous prediction of ice thickness reduction. Use of empirical methods is then the only option. Bilello (1980) analyzed numerous ice measurements of river ice thickness in Alaska and Northern Canada, and proposed the following relationship:

$$h = h_0 - \varepsilon S_5 \quad (2)$$

in which h is expressed in centimeters; h_0 = ice thickness just prior to the beginning of thaw in centimetres; S_5 = accumulated degree-days of “thaw” with respect to a base of -5°C ,

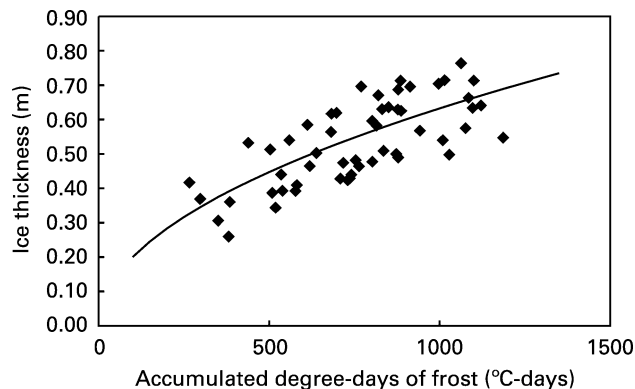


Figure 4 Dependence of ice thickness on accumulated degree-days of frost

which was recommended for river ice conditions; and ε = site-specific empirical coefficient, varying between 0.4 and 1.0 cm/°C-day in the case of Bilello's data.

Prowse *et al.* (1989) found that the lower limit of this range applied to a protracted breakup event in a more southerly maritime river. Beltaos and Burrell (1992) reported values in the range 0.1–0.25 cm/°C-day, based on four years' observations on the Restigouche River, also a maritime stream. Beltaos (1987) found ε to be 0.36 cm/°C-day, on the average, in the lower Thames River in southwestern Ontario, although its value tended to vary from year to year (range = 0.26–0.43). From limited gauge archives, Beltaos (unpublished) calculated ε = 0.28 cm/°C-day for the Saint John River near Fort Kent.

Data for the Blackville site are plotted in Figure 5, and relate to years where the discharge measurements extended beyond the time of maximum thickness (1926, 1930, 1964, 1965, 1973, 1977, 1978, 1982, 1985, 1990 and 1997). The straight line in Figure 5 is a best-fit relationship, after ignoring the low data point at the lower right end of the graph. An obvious discrepancy from the Bilello relationship (Eq. (2)) is that the decay does not start until S_5 attains a value of about 20°C-days. The rate of decay (0.23 cm/°C-day) is considerably lower than Bilello's range, but compatible with previously quoted values in maritime rivers of Eastern Canada.

Onset of breakup

The hydrometric stage records also provide clues as to the timing of the breakup event (Beltaos *et al.* 1990). Of particular interest is the “onset” of breakup, defined as the first time when the winter ice cover is set in sustained motion. To predict the time of breakup initiation, a number of onset criteria have been formulated in the past few decades. Most are completely empirical, relying on various combinations of water level, ice thickness, freeze-up conditions, and air temperature indices such as degree-days of thaw. A common empirical criterion is (Beltaos 1997):

$$H_B - H_F = kh_o - F(S_5) \quad (3)$$

in which H_B = water surface elevation at which the ice cover starts to move; H_F = water surface elevation at which the ice cover formed during the preceding freeze-up event = *freeze-up level*; F = a site-specific function of S_5 , and k = site-specific coefficient, so far known to vary between 2 and 10 – the higher values being associated with relatively tranquil streams and steep banks. Similarly, the function F changes from site to site and from river to river; however, it is always such that $F(0) = 0$. This condition represents “premature” breakup events, characterized by rapid runoff and breakup with minimal, if any,

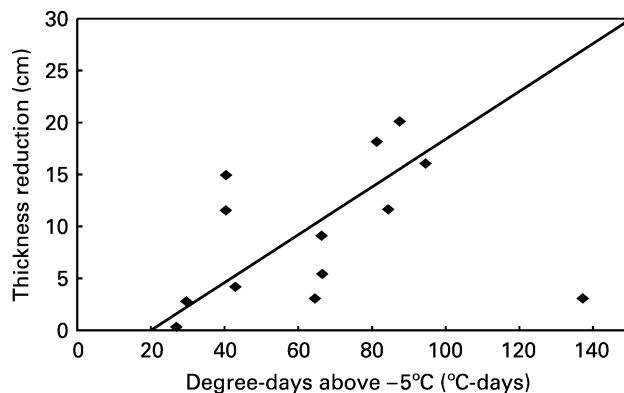


Figure 5 Ice thickness decay prior to breakup

thermal deterioration of the ice cover. Eq. (3) indicates that premature breakup events are initiated when the water level rises above the freeze-up stage by an amount proportional to the ice thickness. Note that this type of criterion does not apply to thermal breakup events, characterized by *in situ* disintegration of the ice cover and insignificant ice breaking or jamming.

For the Blackville site, the available data are not sufficient for direct determination of k , because there are no data points with zero, or near-zero, values of S_5 . Backward extrapolation based on curve-fitting indicates that $k \approx 5.1$.

Eq. (3) and others like it do not explicitly account for hydrodynamic or morphological effects; hence, they can only be applied to the particular river site at which they have been calibrated, i.e. they are *site-specific*. Application to another site on the same river, or to a different river, can only be made if adequate local data are available for calibration. This limitation can, in principle, be overcome with criteria that are based on physical-process hypotheses. A number of these have been proposed in the literature, and were recently reviewed and evaluated using field data from six different river sites (Beltaos 1997, 1999). The following equation, based on the simple requirement that ice plates formed by transverse cracking are set in motion when there is adequate water surface width, was found to adequately describe all six data sets:

$$\Phi_B \equiv \frac{8(W_B - W_i)\varpi_i m^2}{(m - 0.50)h_o} = \beta\sigma_{f_0} \left(\frac{\sigma_f h}{\sigma_{f_0} h_o} \right) = \beta\sigma_{f_0} f(S_5) \quad (4)$$

in which Φ_B represents the dimensionless multi-variable quantity on the left-hand side of the equation; W_B = water surface width at the stage at which the breakup is initiated; W_i = width of ice cover = river width at the freeze-up stage minus the side strips caused by hinge cracking prior to breakup; h , σ_f = ice cover thickness and flexural strength, while the suffix o denotes initial values, just before thermal deterioration begins; m = radius of channel curvature divided by ice cover width; ϖ_i = downslope force per unit area applied on the ice cover by its own weight and by flow shear; and β = dimensionless coefficient between 0.3 and 1.5. The ratio $\sigma_f h / \sigma_{f_0} h_o$ quantifies the loss of ice “competence” due to thermal deterioration during the pre-breakup period. This process involves reductions in both ice thickness via top and bottom melt, and in strength, due to penetrating solar radiation and preferential melting at crystal boundaries (Bulatov 1972; Ashton 1985; Prowse *et al.* 1990). It is difficult to predict such effects, however, owing to complexities introduced by the snow cover and its changing reflective/absorptive properties as melt progresses (Prowse and Marsh 1989). Consequently, the “competence” ratio has been expressed as an empirical function of S_5 .

Eq. (4) is also limited to mechanical breakup events, and can be shown to approximately reduce to Eq. (3) when the site of application is fixed. Although the physical concept expressed by Eq. (4) is rather simple and does not encompass all the details of the initiation of breakup, it explains several trends known by experience. (Note that the difference in widths, $W_B - W_i$, is roughly proportional to the water-level rise, $H_B - H_F$, since W_i is usually close to W_F , and the cross-sectional shape of a river is close to trapezoidal.) For instance, the predicted effect of the freeze-up stage is intuitively plausible and has already been verified in many empirical studies. The effect of river planform (the horizontal geometry of the river, as it is manifested on a topographic map, for instance) is expressed by the dimensionless radius of curvature, m . For relatively straight reaches, m is in the region of 10 or more, and the quantity $(m - 0.50)/m^2$ is about 0.1 or less; for a sharp bend, with $m = 3$, it increases threefold to 0.3. Therefore, relatively straight reaches are expected to break up first and jamming is very likely to occur at sharp bends, as is also known by experience. Similarly,

ice jamming is known to occur where river slope and velocity decrease abruptly, which is explained by low ϖ_i values in Eq. (4).

Data from the Blackville site are plotted in the form suggested by Eq. (4) in Figure 6, together with previous results, including four data points describing observations on the Mackenzie River (Hicks *et al.* 1995). Despite the scatter, the relative collapse of the points, representing 8 rivers of wide-ranging magnitude and slope, provides strong support for the structure of Eq. (4).

Peak breakup stages

The peak stages occurring during the breakup process are usually caused by nearby ice jams that may fully or partly affect the site of interest. The full effect is manifested as the water level associated with an “equilibrium” jam (Beltaos 1995) that extends upstream and downstream of the site. If the jam head (upstream end) is located downstream of the site, the effect is reduced because of the intervening open water reach. If the jam toe (downstream end) is upstream of the site, the effect is again partial and is “felt” when the jam releases, producing a positive flood wave at downstream locations.

Consequently, it may be expected that plots of the maximum breakup stage, H_m , versus the concurrent discharge, Q_m , would be bounded by the equilibrium-jam stage. This is confirmed in Figure 7, where the vast majority of the data points fall below the jam curve. However, there are three points, located in the low-flow range ($< 200 \text{ m}^3/\text{s}$), that lie substantially above the curve. Closer examination of the data revealed that this is caused by the very high hydraulic resistance that applies to low flows, as mentioned in a previous section (“Description of Study Reach and Local Ice Processes”). The equilibrium-jam curve, having been calibrated with relatively large flows, does not take this effect into account. Because the low-flow range is of trivial significance in this application, no attempt has been made to adjust the curve for the higher resistance in this range.

The highest data point in Figure 7 represents the breakup of 1970, which was a premature, mid-winter event. Following a rather cold January, the air temperature rose to -1.4 , 7.8 , and 8.3°C on 2, 3, and 4 February, respectively. This was accompanied by intense rainfall totaling 113 mm, with most of it occurring on 3 February (81 mm). The rise in stage was so rapid that it is not possible to identify the onset of breakup. The peak stage of 8.4 m occurred at 22:00 h on 4 February; by the next day, the mean temperature had dropped to -9.2°C . This event was the most damaging in recent memory: in addition to widespread flooding,

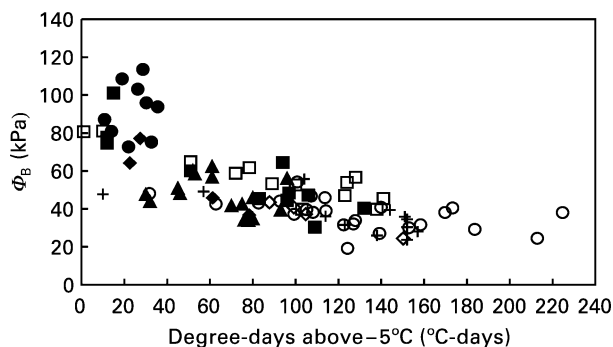


Figure 6 Onset of breakup criterion, tested at 8 sites. After Beltaos (1997), with changes. Hollow circles: Southwest Miramichi River; crosses: Restigouche River; solid diamonds: Grand R.; solid squares: Nashwaak R.; solid triangles: Moose R.; solid circles: Thames R.; hollow squares: Saint John R. (Beltaos 2002); hollow diamonds: Mackenzie R. (data from Hicks *et al.* 1995)

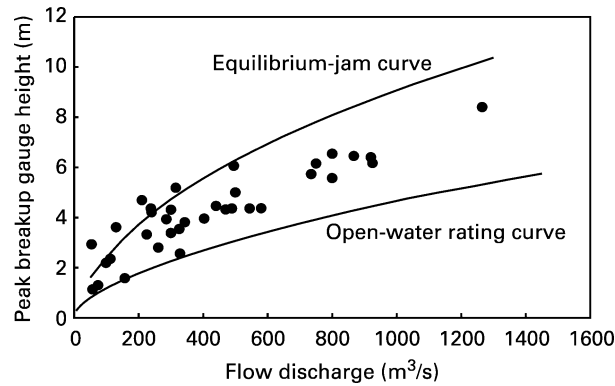


Figure 7 Peak breakup stage as a function of river flow, and effectiveness of the equilibrium-jam curve as an upper boundary of the data points

three bridges, including the one at Blackville, were destroyed, and one fatality was indirectly caused by the flood (Beltaos *et al.* 2001).

In most Canadian rivers, ice-jam floods are just as, or more, severe than open-water ones, despite the fact that the latter are attended by much higher flows. For Blackville, the available data suggest about equal flooding potential in the high-stage range, as illustrated in Figure 8. In this figure, the peak open-water stages have been calculated by means of the gauge rating table, using readily available peak-flow values (HYDAT CD) for different years. In some years, instantaneous flow peaks were unavailable. Corresponding daily-flow peaks were then multiplied by the average instantaneous-to-daily peak ratio (1.11), as determined from those years where both values were available.

It is important to recognize that equality of flooding potential between ice-affected and open-water peaks does not imply that the overall flood frequency is controlled by the latter. The “combined” return periods of high stages (stage exceedance by either process) can be expressed as

$$T_C = \frac{T_i T_o}{T_i + T_o - 1} \quad (5)$$

in which the subscripts C, i, and o denote combined, ice-related, and open-water conditions. With $T_o \approx T_i$, Eq. (5) gives $T_C \approx T_o/2$, which shows that the return period of a given high

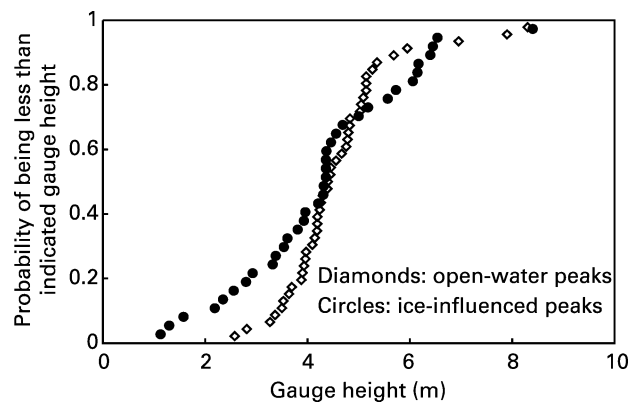


Figure 8 Probability of peak stage occurrence for open-water and ice-covered conditions

stage at Blackville is halved when the ice effect is taken into account. On the other hand, where ice jams dominate the high stages, as for example is the case of the Saint John River near Perth–Andover (Beltaos and Burrell 2002), $T_o \gg T_i$ and Eq. (5) results in $T_C \approx T_i$.

Trends in ice processes

Freeze-up date

As in previous work, the time of freeze-up is defined as the day when a complete ice cover is first established at the gauge site. This definition allows relatively easy identification of the freeze-up date using daily-stage records: it is the day when the stage attains a maximum value following an abrupt rise caused by increasing backwater effects as the downstream ice cover propagates in the upstream direction. Figure 9 summarizes freeze-up dates and hints at a slight trend toward later occurrence, but regression analysis has indicated that this trend is not statistically significant ($P_F = 0.35$; the quantity P_F is the significance level of the F -statistic, i.e. the probability that the regression-derived slope of a trend line could have occurred by chance, if the data belonged to a set with no temporal trend – “null hypothesis”). The freeze-up stage (H_F) showed a slightly negative but not statistically significant trend with time.

Winter ice thickness

To examine whether any trends in the winter ice-cover thickness exist over the years of the record, the available data have been used to either interpolate or extrapolate to 15 March of each year – this being the time when the thickness attains its maximum value or is near the maximum value. These results go back to 1962, and there is also a set for the years 1924–1930, as depicted in Figure 10. The gap between these two data sets is rather large, hence it is not possible to derive a credible trend for both. The post-1961 data points exhibit a slightly negative, but statistically insignificant, trend. On the other hand, Huntington *et al.* (2003) reported a 25 cm decrease in the ice thickness of a central Maine river over the period 1912–2001. The difference accentuates both the strong climatic variability across North-Atlantic regions of the continent and the sensitivity of river ice processes to climatic inputs.

Breakup date

The date of “breakup” in rivers is sometimes taken to be the last day when “backwater” caused by ice is experienced at the gauge site. Backwater is the difference between actual stage and the stage that would have occurred with the same discharge under open-water conditions. This definition allows convenient determination of the breakup date,

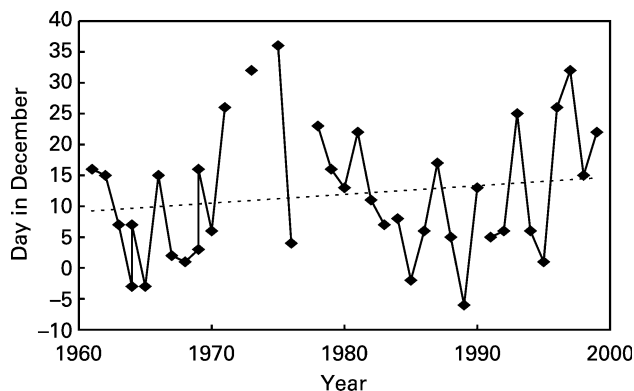


Figure 9 Freeze-up dates during the period of record

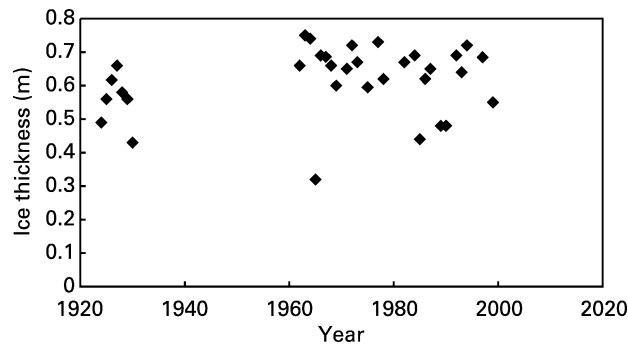


Figure 10 Ice thickness on 15 March during the period of record

i.e. by simply inspecting published daily flow data (Water Survey of Canada), because backwater conditions are designated with the letter “B”. However, it is generally impossible to directly determine the backwater while breakup is in progress, because local access for discharge measurement is, as a rule, unsafe. Designation of the B-condition is then based on subjective judgment and interpolation between dates of site inspections. Another shortcoming of this definition of breakup timing is that the event can be completed well before the last day of “B”, as minor backwater effects can linger as a result of ice remnants in secondary channels and on banks and shallows.

Herein, breakup timing is indexed by the date when breakup is initiated (date of H_B occurrence) and is plotted against the year of occurrence in Figure 11. Although there appears to be a slight trend towards earlier breakups, the trend is not highly significant ($P_F = 0.13$).

Ice jam severity

As already discussed, the severity of ice jams forming near the gauge site is manifested in the peak breakup stage. Figure 12 suggests that the severity of ice jams has increased significantly ($P_F = 0.019$) in the past 40 years or so, causing a stage increase of over 1 m, on the average.

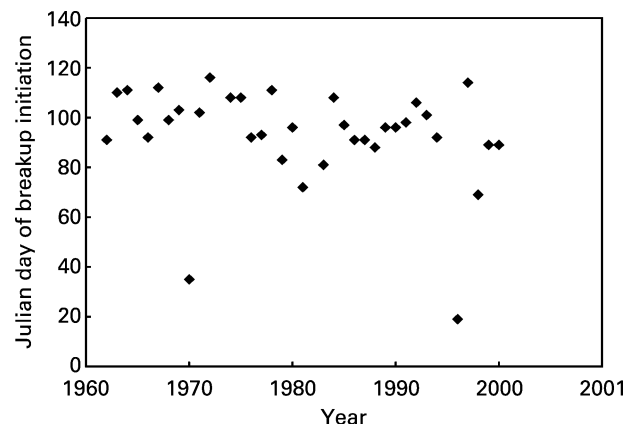


Figure 11 Breakup dates during the period of record

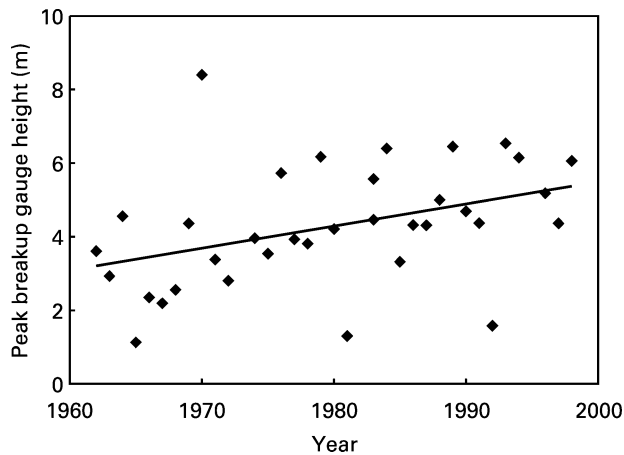


Figure 12 Peak breakup stages during the period of record. The linear-regression fit indicates a positive trend

Trends in hydrologic variables

The flow hydrograph is a major factor in the evolution of river ice processes, particularly during the fall and spring periods. Consequently, trends in flows may help to explain some of the trends in ice processes.

For instance, it is known that the higher the flow in the river the more prolonged the freeze-up process, because of the tendency to form thicker covers at higher flows. Since freeze-up mostly occurs in December, the mean December flow was examined and found to exhibit a slight, and statistically insignificant, positive trend. This is in line with the lack of a significant trend in freeze-up dates.

Ice-jam severity is also related to flow, as has been illustrated in Figure 7. Here, however, the most pertinent variable would be the maximum daily flow in April, since breakup typically occurs in that month and the peak breakup stage is governed by runoff events. Figure 13 shows that there is a trend for maximum April flows to increase over the period of record, in accord with the trend of Figure 7. For river flows, conventional parametric tests of trend significance, such as the F-test, may not be meaningful, because the regression residuals themselves exhibit temporal trends and are not normally distributed. The frequently used Mann–Kendall test (Hirsch *et al.* 1982; Burn 1994), which is based on the signs of

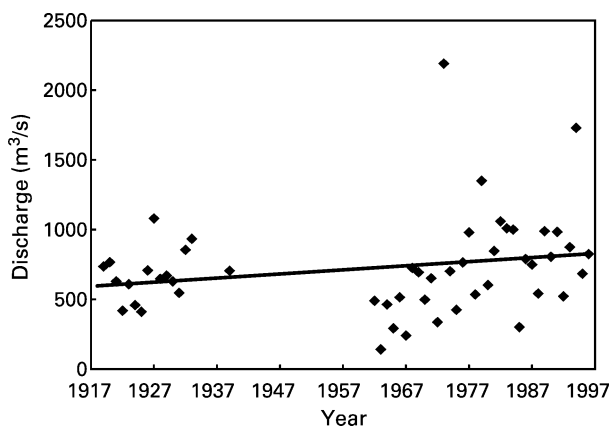


Figure 13 Maximum daily flow in April during the period of record

differences between temporally shifted pairs of a variable, is adopted herein for flow, and later on, for rainfall. The significance probability associated with the trend shown in Figure 14 is $P_{MK} = 0.026$.

Two data points in Figure 14 are associated with unusually high flows (1973, 1994), and it is of interest to consider how they affect the overall trend. A trend analysis on a screened data set (i.e. after removing the 1973 and 1994 data) indicated that the trend slope remains positive but is halved. The significance probability increases to 0.095, suggesting that the trend is still significant, but less so than in the unscreened data set.

As indicated in Table 1, significant positive trends have also been found for the maximum daily flow in January, February and March, while a negative trend for May was not statistically significant. The increasing winter flows suggest that mid-winter breakup events may become more frequent in the future, as is already occurring in the upper Saint John River (Beltaos 2002).

Climatic conditions

Data sources

Weather records (to 1996) for Eastern Canada can be purchased from Environment Canada in the form of a CD. The most representative meteorological station for the present study is at Doaktown, located some 45 km SW of Blackville, but its records only go back to 1934. These can be supplemented, however, by the archives of a station at Chatham (now a part of the City of Miramichi), which extend from 1873 to 1947. Chatham is located about 45 km NE of Blackville. More detailed information on these two stations is presented in Table 2.

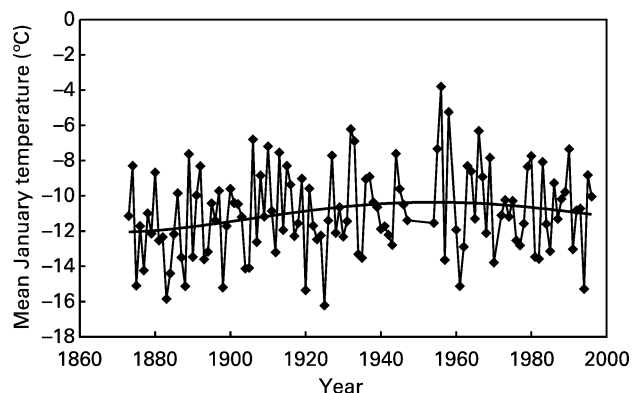


Figure 14 Mean January air temperature, from records at Chatham (1873–1947) and Doaktown (1952–1996)

Table 1 Trend-analysis results for maximum daily discharge

| Month | Years of record | Linear-fit slope ($\text{m}^3 \text{s}^{-1} \text{ year}$) | P_{MK} | Data not available for: |
|----------|-----------------|--|----------|-------------------------|
| January | 1919–1996 | 1.10 | 0.074 | 1934–1938, 1940–1961 |
| February | 1919–1996 | 1.20 | 0.026 | 1934–1938, 1940–1961 |
| March | 1919–1996 | 0.98 | 0.092 | 1934–1938, 1940–1961 |
| April | 1919–1996 | 2.94 | 0.026 | 1934–1938, 1940–1961 |
| May | 1919–1996 | -2.27 | 0.472 | 1934–1938, 1940–1961 |
| November | 1918–1996 | 0.80 | 0.304 | 1934–1937, 1939–1960 |
| December | 1918–1996 | 1.60 | 0.442 | 1934–1937, 1939–1960 |

Table 2 Climatic data stations near Blackville that have been used in this study

| Station name and number | Latitude (N) | Longitude (W) | Elevation (m) | Years of record |
|-------------------------|--------------|---------------|---------------|--|
| Doaktown 8101200 | 46°33' | 66°09' | 38 m | Temperature: 1952–1996 Precipitation: 1934–1996 |
| Chatham 8100990 | 47°03' | 65°29' | 30 m | Temperature: 1873–1947 Precipitation: 1873–1947 |
| Miramichi Airport | 47°01' | 65°27' | 34 m | Temperature: 1943–1996 Precipitation: 1943–1996 |

Due to gaps in the temperature record at Chatham, there is little overlap with the Doaktown record to enable direct comparisons of concurrent temperature values at the two stations. This difficulty was resolved by considering a third nearby station, at Miramichi Airport, which is located very near the Chatham station and overlaps with both the Chatham and Doaktown records (Table 2). Comparisons between Miramichi and Chatham and Doaktown and Miramichi have indicated near-coincidence of concurrent monthly mean air temperatures (linear-fit slope, $R^2 = 1.04$, 0.979 for Doaktown to Miramichi; and 0.975, 0.996 for Miramichi to Chatham; the implied relationship between Doaktown and Chatham is $T_{\text{Doaktown}} = 1.014T_{\text{Chatham}} - 0.12$; T = monthly mean temperature in °C).

Air temperature

Both ice processes and flow hydrographs are influenced by climate, especially air temperature and precipitation. Mean monthly temperatures at the first two stations of Table 2 have been plotted against year of occurrence, as illustrated in Figure 14. The polynomial fit of the data points shows that a warming trend to about the 1950s is succeeded by cooling. The cooling trend is consistent with findings by Brimley and Freeman (1997) on many Atlantic meteorological stations for the past 40 years or so, especially in Nova Scotia and Newfoundland. These authors noted that the winter temperatures off the coasts of these provinces have been dropping since the late 1960s, and suggested a possible link to the North Atlantic Oscillation (NAO).

Table 3 summarizes statistical tests performed on mean monthly air temperature for the Doaktown meteorological station, whose record represents the second half of the 20th century. Late fall and winter months exhibit cooling trends (although not always statistically significant), while spring, summer, and early fall months appear to be warming.

Table 3 Trend-analysis results for monthly mean air temperature at Doaktown

| Month | Years of record | Linear-fit slope (°C/century) | P_F | Data not available for: |
|-----------|-----------------|-------------------------------|--------|-------------------------|
| January | 1954–1996 | –3.7 | 0.26 | 1959, 1971 |
| February | 1954–1996 | –1.5 | 0.60 | 1962 |
| March | 1953–1996 | 0.6 | 0.81 | 1959 |
| April | 1953–1996 | 2.2 | 0.12 | 1960 |
| May | 1953–1996 | 4.6 | 0.016 | 1954, 1961 |
| June | 1953–1996 | 5.8 | 0.0004 | 1954 |
| July | 1953–1996 | 3.8 | 0.045 | 1954, 1960 |
| August | 1952–1996 | 5.2 | 0.0003 | 1954, 1960 |
| September | 1952–1996 | 2.1 | 0.14 | 1954, 1958 |
| October | 1953–1996 | 0.2 | 0.87 | 1954, 1958, 1960 |
| November | 1952–1996 | –4.1 | 0.056 | 1954, 1958, 1960, 1961 |
| December | 1953–1996 | –0.4 | 0.91 | 1954, 1958, 1961 |

As found by Beltaos (2002) an important index of changes in winter hydrology is the total number of “mild” days during the winter months, owing to its connection with the probability of rainfall occurring during the same period. A mild day is defined as one with mean daily air temperature above 0 °C. In the case of the upper Saint John River in NW New Brunswick, the number of mild days has increased in the past 80 years or so, in step with slight winter warming. A similar trend ($P_{MK} = 0.016$) is apparent in the present case (Figure 15), despite the cooling trends in the corresponding mean monthly temperatures.

Precipitation

Considering first rainfall, which is a major driving factor of the breakup and jamming processes, Table 4 indicates positive trends for winter and early spring (January to April) for the period 1940–1996. All but February’s trend are statistically significant, and are consistent with the corresponding positive trends in flows indicated in Table 1, and with the increasing number of mild winter days illustrated in Figure 15. Similar trends have been found for monthly snowfall amounts (Table 4), while the accumulated snowfall during 1 December to 31 March has a positive trend of 86 cm/century ($P_{MK} = 0.10$).

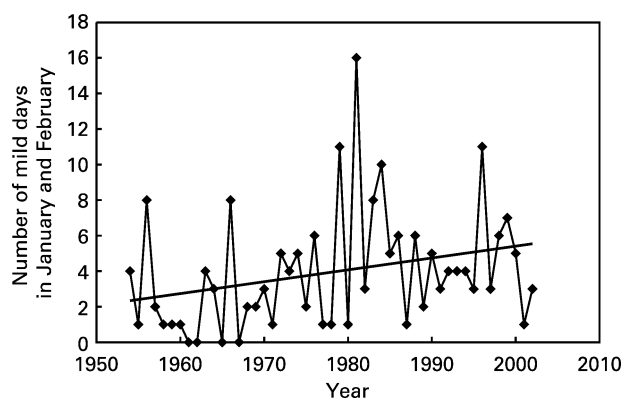


Figure 15 Number of mild days during January and February at Doaktown

Table 4 Trend-analysis results for monthly precipitation at Doaktown

| Month | Years of record | Rainfall linear-fit slope (mm/year) | P_{MK} | Data not available for: | Snowfall trend |
|-----------|-----------------|-------------------------------------|----------|-------------------------|-----------------|
| January | 1940–1996 | 0.38 | 0.018 | 1942–1944, 1963 | Positive |
| February | 1940–1996 | 0.24 | 0.312 | 1942, 43, 1960 | Negative |
| March | 1940–1996 | 0.28 | 0.066 | 1942, 1943 | Positive |
| April | 1940–1996 | 0.50 | 0.024 | 1942, 1943 | No trend |
| May | 1940–1996 | 1.30 | 0.0006 | 1942, 1943 | NE ¹ |
| June | 1934–1996 | 0.59 | 0.168 | 1935 | NE |
| July | 1934–1996 | 0.63 | 0.020 | 1935 | NE |
| August | 1934–1996 | 0.56 | 0.078 | 1935 | NE |
| September | 1934–1996 | 0.15 | 0.436 | 1935 | NE |
| October | 1934–1996 | 0.76 | 0.026 | 1935–1937, 1942–1944 | NE |
| November | 1939–1996 | 0.45 | 0.106 | 1942, 1943, 1968 | No trend |
| December | 1939–1996 | 0.31 | 0.358 | 1942, 1943 | Positive |

¹ Not examined

Discussion

The main facets of the ice regime, i.e. timing of freeze-up and breakup, ice growth and decay, onset of breakup, and ice-jam water levels, have been documented for the Southwest Miramichi river in the previous sections, using hydrometric station records. Over the length of the record, the river's ice processes exhibit several trends, but there are also instances where no significant trend could be detected. As outlined in this section, the present findings are consistent with concurrent variations, or lack of, in related hydroclimatic factors.

Considering first the timing of freeze-up, the present data have indicated no significant trend for the past 40 years or so. This is consistent with the lack of significant trends in December flows and temperatures, since flow and air temperature are the main factors that govern freeze-up at a fixed location (Andres and Van der Vinne 1998).

Winter ice growth adheres to the Stefan equation, but seems to be faster than commonly recommended values of growth coefficients. The slight cooling detected for the winter months does not seem to have influenced the 15 March ice cover thickness. Application of Eq. (1) with estimated representative values of the accumulated degree-days of frost for the 1950s and the 1990s has indicated that the difference in ice thickness would have only been about 2 cm. This would be too low to enable trend detection over the same period.

Ice thickness decay in the spring does not agree closely with the Billello equation, suggesting that other factors, in addition to accumulated degree-days of thaw, are at work. On the other hand, the onset of breakup occurs in accordance with Eq. (4), in agreement with 7 other river sites. This underscores the transferability of physics-based predictions under different morphologic and climatic conditions. A slight trend toward earlier breakups is consistent with higher winter and early spring flows and increasing April temperatures.

Ice jamming is manifested in peak breakup stages that are well above the open-water rating curve but, with few exceptions, fall under the theoretical upper bound curve that describes equilibrium ice jams in the gauge reach. A comparative frequency analysis indicated that ice jams are about as likely to cause high water levels as are open-water floods. This indicates that the potential for ice-jam damage is considerable, although not as severe as in the Saint John, where ice jams completely dominate the major-flood frequencies (Beltaos and Burrell 2002). The trend toward increasing peak breakup stages is consistent with increasing April flow and rainfall and possibly with the increasing winter snowfall.

Increasing ice-jam severity is also a concern with respect to infrastructure, which typically has a design life of 50 years or more. There is no methodology at present to account for higher and more frequent flood stages in the future. Simple linear extrapolation of current trends is not adequate, because warming is expected to accelerate in the next 50–100 years. The rational approach would be to utilize a “cascade” of models, such as climatic, hydrologic, hydraulic, and river-ice models. However, gaps in knowledge throughout the “cascade” are likely to produce serious uncertainties in final predictions. Nevertheless, this approach would be useful for scenario assessment and sensitivity analysis. Its early implementation and repeated use would also help resolve various practical problems that will invariably arise from differences in the spatial and temporal scales utilized by the various models.

Despite the slight winter cooling in the past 40 years or so, the number of mild days occurring in January and February has increased, possibly as a result of increased day-to-day variability. This enhances the probability that a snowstorm may be transformed to a rainfall event, and is consistent with the increased monthly rainfall and flow discharge. If these trends continue, major winter breakup events, such the one of February 1970, will become more frequent. Not only is this a socio-economic concern, but it could have serious impacts on aquatic life that is adapted to tranquil winter conditions, as has been found by Cunjak *et al.* (1998).

Conclusions

An ice cover typically forms on the Southwest Miramichi River near Blackville in December and breaks up in April. As is often the case in Atlantic Canada, mid-winter thaws are relatively frequent, but do not often produce sufficient runoff to cause breakup of the ice cover. When they do, devastating floods can ensue, such as the flood of February 1970.

Trends in ice processes that have been detected over a 40-year period are consistent with concurrent trends in climatic variables and with consequent trends in the flow hydrograph. Of these, the most important is the increase in peak water levels during the ice breakup, signaling an increasing ice-jam severity. Increasing winter flows are not yet capable of producing more frequent winter breakups, but this may change in the future, as warming effects accelerate.

Consistent with other findings in parts of Atlantic Canada, winter temperatures are cooling slightly; however, winter rainfall and snowfall are on the increase. The increase in rainfall is consistent with an increase in the incidence of mild winter days. The latter finding, however, is in the opposite direction to the cooling trends in monthly temperatures and points to increasing day-to-day variability.

Quantitative descriptions of ice processes in the Southwest Miramichi river, such as Eqs (1)–(4) and the “equilibrium-jam” theory, are in general accord with findings in other rivers, although some of the relevant coefficients differ. Ice jams do not dominate the frequency of local floods, as they often do in Canadian rivers, but still play a large part.

Acknowledgments

This study was partially motivated by an ongoing research program on the interactions between river ice and bridges, which is carried out in collaboration with the New Brunswick Departments of Transportation and Environment and Local Government. The study also forms a part of a national research program carried out by NWRI to study climate impacts on extreme ice-jam events. Extensive hydrometric station archives were kindly provided by Paul Noseworthy, Hydrometric Supervisor of the Fredericton Environmental Monitoring office of Environment Canada. Much of the data processing and preliminary analysis were carried out by Steve Howell under special contract.

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