Comparing the impacts of regulation and climate on ice-jam flooding of the Peace-Athabasca Delta

Spyros Beltaos

National Water Research Institute, Watershed Hydrology and Ecology Research Division, Environment Canada, 867 Lakeshore Road, Burlington, ON L7R 4A6, Canada

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ABSTRACT

The Peace-Athabasca Delta (PAD) in northern Alberta is one of the world’s largest inland freshwater deltas, home to large populations of waterfowl, muskrat, beaver, and free-ranging wood bison. Following construction of the W.A.C. Bennett Dam in the late 1960s, a paucity of ice-jam flooding in the lower Peace River has resulted in prolonged dry periods and considerable reduction in the area covered by lakes and ponds that provide habitat for aquatic life in the PAD region. Past studies have shown that both regulation and climate have contributed to the drying trend, via increased freezeup levels and reduced breakup flows, respectively. However, it has not so far been possible to assess the relative impacts of these two factors. This question is addressed herein by developing quantitative relationships based on the concept of conditional probability. The new methodology is first tested against observed frequencies of ice-jam floods both before and after regulation; it is subsequently applied to quantify the effects of climate and of regulation on the post-regulation reduction in ice-jam flood frequency. The results indicate that both factors have contributed significantly to the drying of the PAD, with regulation having had the more pronounced effect. Complications that may arise from secondary effects of regulation on breakup flows or of climate on freezeup levels are discussed and quantified. It is shown that such effects do not alter the overall conclusions of the paper.

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1. Introduction

The Peace-Athabasca Delta (PAD) in northern Alberta (Fig. 1) is one of the world’s largest inland freshwater Deltas, home to large populations of waterfowl, muskrat, beaver, and free-ranging wood bison. It was protected in 1922 with the creation of Wood Buffalo National Park, one of Canada’s 15 UNESCO World Heritage Sites. In 1982, the PAD was designated as a Ramsar Wetland of International Importance in recognition of its ecological, historical, and cultural value. To the present day, the PAD continues to support traditional fishing, hunting and trapping activities.

After the construction of the W.A.C. Bennett Dam for hydropower generation during the period 1968 to 1971, this complex and dynamic region has experienced prolonged dry periods, and considerable reduction in the area covered by lakes and ponds that provide habitat for aquatic life. Previous work has suggested that the combined effects of flow regulation and climatic variation have inhibited the formation of extensive spring ice jams in the lower reach of Peace River that trigger much of the Delta inundation (Beltaos et al., 2006; Prowse and Conly, 1998). Ice jams are known to cause much higher water levels than open water floods and are particularly effective in replenishing the higher-elevation, or “perched”, basins of the PAD.

The effect of climate is primarily manifested in decreasing flows during the spring breakup while that of regulation is mainly exerted via increased freezeup levels (Andres, 1996), which in turn reduce the chances for ice-jam formation. An important question that arises from the scientific findings to date is how to quantify and compare the contributions of climate and regulation to the drying of the PAD. The answer to this question would not only advance current knowledge with respect to the specific issue of how the regulation of Peace River, by means of structures located some 1100 km upstream, has affected the ecology of the PAD. It would also enhance current capability to anticipate and quantify the effects of regulation in other ice-laden rivers. The need for this kind of knowledge is particularly acute under conditions of a changing climate, and especially in northern regions of the globe where climatic change is the most pronounced.

The objective of this paper is twofold: (a) develop quantitative methodology for delineating simultaneously-operating effects of regulation and climate on the frequency of ice-jam flooding; and (b) apply this methodology to the lower Peace River in order to assess and compare these effects in the drying of the perched basins of the PAD.

Following a discussion of background information on ice breakup processes in general and in the lower Peace River in particular, various data sources are identified. Next, the effects of regulation and of climate on ice-jam frequency are quantified via development of mathematical equations based on conditional probabilities. These relationships are validated using observed flood frequencies during the pre- and post-

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regulation periods. Subsequently, the probabilistic methodology is applied to calculate and compare the individual contributions of regulation and of climate on the experienced decrease in ice-jam flood frequency. Secondary considerations arising from potential effects of climate on freezeup levels and of regulation on breakup flows are discussed.

2. Background information

The breakup of river ice is triggered by mild weather and encompasses a variety of processes associated with thermal deterioration, initial fracture, dislodgment and mobilization of the ice cover, fragmentation, transport, jamming, and final clearance of the ice. During the pre-breakup phase, the ice cover becomes more susceptible to fracture and movement via thermally induced reductions in its thickness and strength. At the same time, the warming weather brings about increased flow discharges, due to snowmelt and/or rainfall. The rising flows and water levels fracture the ice cover and reduce its attachment to the riverbanks while the increased hydrodynamic forces cause it to move and break down into relatively small blocks. This is the onset of breakup, and is followed by the drive, that is, the transport of ice blocks and slabs by the current. The onset is governed by many factors,

Fig. 1. Plan view of lower Peace River and Peace-Athabasca Delta.

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including channel morphology, which is highly variable along the river. It is thus common to find reaches where breakup has started, alternating with reaches where the winter ice cover has not yet moved.

Depending on hydro-meteorological conditions, a breakup event can be thermal or mechanical. The former type normally occurs when mild weather is accompanied by low runoff due to gradual melt and lack of rain. The ice cover deteriorates in place and eventually disintegrates under the weak forces applied by the modest current. Ice jamming is minimal, if any, and water levels remain low. Mechanical breakup occurs where the runoff is sufficient to lift and dislodge the cover while it retains a sizeable portion of its mechanical strength. This type of event can result in ice jamming of varying severity, depending on hydro-climatic conditions. At the upper end of severity, mechanical events have also been called premature or dynamic. This is the type of event that can lead to large-scale ice jam flooding of the PAD.

In the colder continental parts of Canada, such as the Prairies or Territories, one is most familiar with a single event, the spring breakup, which is typically triggered by snowmelt. This is also the case for the lower Peace River, where rainfall, if any, contributes little to the total spring flow. In more temperate regions, such as parts of Atlantic Canada, Quebec, Ontario and British Columbia, events called midwinter thaws are common. Usually occurring in January and February, they comprise a few days of mild weather and typically come with significant rainfall. River flows may rise very rapidly and sufficiently to trigger a winter breakup on many local rivers, an event that is typical of the dynamic variety.

Ice jams form where moving ice blocks, produced by the breakdown of dislodged segments of the ice cover, encounter relatively intact, stationary ice. As shown by Beltaos (1997), this occurrence is caused by the characteristic irregularity of natural streams, in which morphologically different reaches exhibit different susceptibility to the dislodgement of the winter ice cover. The most obvious ice-jamming sites are sharp bends, abrupt reductions in slope or flow velocity (e.g. river mouths or reservoirs), and channel constrictions. Streamwise variability in ice strength and thickness, as controlled by the nature of the freezeup ice cover types, may also play a role (Prowse et al., 1990). The water levels caused by breakup ice jams are primarily determined by flow discharge, ice volume available to form the jam, and channel slope and width (Beltaos, 1995).

Ice jams can stay in place for a few minutes or for many days; they can be a few hundred meters or many kilometers long. Owing to characteristically large aggregate thickness and underside roughness (Beltaos, 1995), ice jams can cause very high water levels, many meters above the equivalent-discharge, open-water flow stages. When a jam lets go, a large amount of water comes out of storage in short time, producing a “jave” (short for wave generated by the release of an ice jam). The water level drops precipitously upstream of the jam, but rises rapidly downstream; at the same time, water speeds can increase to extreme values while the wave propagates at even higher rates. Intact ice cover may be broken up and carried by the jave; if it is still very competent, it may stay in place and initiate another jam. In this manner, more and more ice is broken up and carried down the river, until the final jam releases. This is the start of the final clearance of ice.

Based on the preceding considerations, it can be generally stated that an ice-jam flood will occur when the following conditions are fulfilled:

(a) The breakup event must be of the mechanical type, else there cannot be significant jamming;
(b) A sizeable ice jam must form and affect water levels at the site of interest, which may or may not happen, even during a mechanical event; and
(c) The flow discharge must exceed a certain minimum value that is governed by local hydraulic parameters and floodplain configuration.

For the Delta reach of Peace River (Sweetgrass Landing to mouth of Peace, Fig. 1), the minimum flow for ice-jam flooding has been estimated as ~4000 m³/s by applying the numerical model RIVJAM to known ice-jamming (toe) locations, such as Moose Island, Rocky Point, mouth of Peace (confluence of R. Des Rochers), and island in the Slave River located ~10 km downstream of the mouth of Peace, as shown in Fig. 1 (Beltaos, 2003a).

3. Ice regime and hydrometric data

There is very limited information concerning the freezeup and breakup processes within the Delta reach of the lower Peace. Historical ice-jam flood events have been reviewed by Demuth et al. (1996), Giroux (1997a, 1997b), Peterson (1994), and Timoney (2009) who cited various sources of historical information. However, for the quantitative analysis that is needed in the present study, the only adequate source is the hydrometric record of the gauge at Peace Point, which is operated by the Water Survey of Canada (WSC). This gauge is located well upstream of the Delta reach (~100 km above the mouth of Peace) or some 50 km above Sweetgrass Landing. However, the local channel configuration and slope are not very different from those encountered within the Delta reach. Hence Peace Point records have been used as a surrogate for Delta-reach processes, with the understanding that various details may not always be adequately represented. This approach was first adopted by Prowse et al. (1996) and later by Beltaos et al. (2006b). Salient features of the Peace Point gauge and of other gauges used in the present study are summarized in Table 1.

Flow data for any one gauge can be readily accessed online and represent daily mean values. As such, they largely reflect river “carrier” flow, i.e. basin spring runoff plus any outflows from regulation structures. Contributions to daily mean values by occasional javes are possible, but would be relatively small because jave influence on flow is highly transient. More detailed information, such as 15-minute and daily mean water levels, as well as ice thickness, can be obtained on request from the appropriate regional office of WSC. For some gauges, daily mean water levels are also accessible online (e.g. Peace Point, after 2001).

Data analysis and interpretation also require knowledge of the hydraulic characteristics of the gauge site, customarily expressed as reach-average bathymetry and slope in a representative reach centered at the site itself. To this end, five cross sections were surveyed in the Fall of 1999, spaced 1 km apart, while the average open-water surface slope in the same reach was measured as 0.000064. [For an 8 km gauge reach, Kellerhals et al. (1972) reported a measured slope value of 0.000074 while other data in their report indicate that the slope decreases towards the river mouth, having an average value of 0.000050 between the latter and Peace Point.] In addition to the relatively short Peace Point segment, many other cross sections have been surveyed by various agencies along the reach extending from Boyer Rapids to ~10 km into the Slave River (Fig. 1). For more details on channel hydraulics see Beltaos (2003a); Beltaos et al. (2006); Demuth et al. (1996).

4. Significance of freezeup level and breakup discharge

The onset of a mechanical event requires the river level to rise above the freezeup level by an amount that depends on local river bathymetry and morphology as well as ice thickness and strength. Herein, freezeup level (Hf) is defined as the water surface elevation at which the ice cover formed during the preceding freezeup event. The influence of Hf on the onset of breakup was first detected empirically and later explained by physical arguments (e.g. Beltaos, 1990, 1997; Shulyakovskyi, 1966, 1972). Essentially, the freezeup level defines the width of the ice cover, which will not move unless there is sufficient room on the water surface to allow mobility and adequate hydraulic forces are applied on it by the flow to set it adrift. This is a simplified qualitative description of the basic physical processes involved (Beltaos, 1997).
For discussion purposes, it is convenient to rearrange Beltaos’ (1997) ice mobilization formula (breakup onset criterion, as follows (Beltaos and Carter, 2009):

\[ \frac{W-W_i}{W_R-W_i} = 0.5 \sigma_s = \frac{h_o}{8m^2} \left( \frac{W_R-W_i}{W-W_i} \right) f_c(\Sigma T_{50}) \]  

(1)

in which the quantity \( \sigma_s \) has units of stress and is proportional to the flexural strength of the ice cover prior to the start of thermal decay (\( \sigma_s \approx 70-120 \) kPa); \( W \) = water surface width at the stage at which the breakup is initiated; \( W_i \) = width of ice cover = river width at the freezeup stage minus side strips caused by hinge cracking prior to breakup; \( W_R \) = channel width at a conveniently selected reference stage for the site under consideration; \( m \) = radius of channel curvature divided by ice cover width; \( \sigma_s \) = ice “tractive stress” = downslope force per unit surface area applied on the ice cover by its own weight and by flow shear; and \( f_c \) is the declining “ice competence” (defined as the product of ice thickness with flexural ice strength), expressed as a ratio relative to its non-decayed value. By definition, \( f_c \) ranges from 0 to 1. Lack of ancillary data and of safe river survey when \( f_c \) is less than 1, makes it very difficult to calculate or measure its value. For practical applications, this quantity has been replaced by a site specific function of the empirical parameter \( \Sigma T_{50} \), the accumulated degree-days of thaw referred to a base temperature of \(-5^\circ C\) (Billelo, 1980).

The left-hand side (LHS) and the right-hand side (RHS) of Eq. (1) may be thought of as the driving stress (D), and the resistance to dislodgement (R), respectively; both have units of force per unit area. The driving stress D is largely governed by the ice competence of the freezeup stage via \( \Sigma T_{50} \) but is small because it appears on both terms of the width-difference ratio. The resisting stress R increases with increasing ice thickness, increasing channel curvature, and increasing freezeup stage. The latter is not immediately obvious, but can be inferred by noting that the higher the freezeup stage, the greater is \( W_R \), hence the lesser is the difference \( W_R-W_i \), and therefore, the greater is the inverse of that difference. At any given time, both D and R vary along the river, because most of the related physical parameters are subject to spatial variability (m, \( W, W_R, W_i, h_o, f_c \)). This is particularly so for the dimensionless radius of curvature, \( m \), which can span more than an order of magnitude. Large values of \( m \) which typify relatively straight reaches, reduce resistance and thence enhance amenability to dislodgement. The converse is also true and explains why sharp bends are prime jamming sites.

In rivers of ordinary slope (order of 0.0001 or greater), the driving stress generated by the rising flow is usually enough to dislodge the less resistant segments of the ice cover. This process leads to a variety of ice conditions along the river, with jams forming where more resistant ice cover remains in place to arrest ice that has been dislodged in upstream reaches and broken down into blocks along the way. There is little order in the manner of breakup progression and there can be several breakup fronts at any one time within a long reach.

On the other hand, very flat rivers may require jave action (via amplified \( \sigma_s \) in Eq. (1)) to initiate breakup, as found by Beltaos (2007) for the Peace Point reach. Near Peace Point, the river slope is well below 0.0001. With the exception of localized rapids sections, similarly low values are representative of the last 570 km of the Peace, extending from near Carcajou, Alberta, to the river mouth at the head of the Slave River (Kellerhals et al., 1972). By contrast, the reach between the town of Peace River and Carcajou has an average slope of 0.00021. Where the breakup is largely driven by javes, the front progression should be generally in the downstream direction, which is in accord with observations. Secondary fronts that sometimes develop ahead of the main front (Friesenhan, 2005) could be caused by spontaneous ice dislodgement in a relatively straight reach, but jave action cannot be discounted either, as outlined next.

When a jave is advancing into an ice covered reach, it will encounter variable resistance to dislodgment. It is possible that a particularly resistant ice segment remains in place while downstream segments are cleared of ice. When the upstream ice run arrives (well after the faster-moving waveform), it will be arrested by the resistant ice cover and form a front with a jam behind it. At the same time, the downstream ice has already been set in motion, and will come to rest against the next resistant ice segment, thus forming a secondary front. An intimation of this process was provided by a May 1, 2003 jave, which is known to have dislodged several ice sheets near Peace Point, even though the main breakup front stalled at Boyer Rapids (Beltaos, 2007).

In jave-driven breakup, the driving stress (Eq. (1)) is largely determined by the backwater created by the “parent” jam. Ice-jam backwater increases with river flow and, up to a maximum value, with jam length (Beltaos, 1995). The distance over which the cover is dislodged by any one jave, and thence the rate of breakup progression, will thus be greater for higher flows and longer jams. In turn, this allows less time for thermal decay of the downstream ice cover, and enhances the probability of a mechanical event. Moreover, breakup initiation is impeded by higher freezeup stage and thicker ice cover (via \( W_i \) and \( h_o \)), respectively, in Eq. (1)). Consequently, thinner ice covers and lower freezeup levels would have qualitatively similar effects on the rate of breakup progression as do higher flows and longer jams, and tend to also promote mechanical events.

For the Peace Point reach, Beltaos and Carter (2009) found that the term \( h_o/(W_R-W_i) \), which expresses the effects of ice thickness and freezeup level on the resistance to dislodgment (Eq. (1)), is very closely approximated by the relationship:

\[ r_d \approx h_o/(W_R-W_i) \approx 0.0023 h_o^{0.5}(H_f-208.8)^{1.5} \]  

(2)

This equation indicates that \( H_f \) is by far the dominant resistance factor, as the exponent of \( h_o \) is relatively small and so is the inter-annual variability of ice thickness. Consequently, high freezeup levels can play a major inhibiting role in the occurrence of mechanical and dynamic breakup events, and thence of ice jams. This is especially applicable

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<table>
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<tr>
<th>Gauge name</th>
<th>Gauge number</th>
<th>Years of record</th>
<th>Latitude (°N) and longitude (°W)</th>
<th>River distance (km) below Bennett Dam(1)</th>
<th>Gross drainage area (km²)</th>
</tr>
</thead>
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<tr>
<td>Peace River at Hudson Hope</td>
<td>07EJ001</td>
<td>1917–2011</td>
<td>56°1’39” 121°5’36”</td>
<td>30</td>
<td>73100</td>
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<tr>
<td>Peace River at Peace River</td>
<td>07HA001</td>
<td>1915–2011</td>
<td>56°14’41” 117°18’51”</td>
<td>410</td>
<td>194374</td>
</tr>
<tr>
<td>Peace River at Peace Point</td>
<td>07KC001</td>
<td>1959–2012</td>
<td>59°7’5” 112°26’13”</td>
<td>1150</td>
<td>293000</td>
</tr>
<tr>
<td>Smoky River at Watino</td>
<td>07CJ001</td>
<td>1915–2011</td>
<td>55°42’52” 117°37’23”</td>
<td>402 (2)</td>
<td>50300</td>
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<tr>
<td>Wabasca R. at highway No.88</td>
<td>07ID002</td>
<td>1970–2011</td>
<td>57°52’28” 115°23’20”</td>
<td>895 (2)</td>
<td>35800</td>
</tr>
</tbody>
</table>

(1) Based on data provided in Kellerhals et al. (1972).
(2) Tributary mouth.

Table 1

Locations of Water Survey of Canada (WSC) gauges used in present study.
where the river is regulated for hydropower generation because fall and winter flows are much larger than they were in their natural state. These higher flows cause higher-than-natural freezeup levels, which in turn require larger spring flows to trigger mechanical breakup events. A good example is the 2003 breakup in lower Peace River. In situ observations indicated that the event was mechanical with consecutive ice jams and ice runs between the town of Peace River and Peace Point, but reverted to thermal mode downstream of Peace Point (Beltaos, 2007). As shown in Table 2, the maximum flow ($Q_{\text{max}}$) attained during the breakup event is comparable to the maximum breakup flows associated with the ice-jam floods of 1972 and 1996. Yet, the 2003 flow was not even sufficient to dislodge and mobilize the ice cover, let alone to cause flooding, beyond Peace Point. This is explained by the respective values of the resistance parameter, $r_d$: for 2003, $r_d = 0.041$, which is more than twice the value for 1972 and almost 3 times that for 1996.

Relevant hydro-climatic studies have concluded that the effect of increased freezeup levels has been partly responsible for the post-regulation drying trend of the perched basins of the PAD (Beltaos, 2003b; Beltaos et al., 2006; Prowse and Conly, 1998; Prowse et al., 1996). These researchers also found an adverse climatic influence via reduced breakup flows, especially those of the unregulated Smoky River, the largest tributary of the Peace.

The positive effect of breakup discharge and the inhibiting effect of freezeup stage on ice-jam flood occurrence are aptly illustrated in Fig. 2, which shows that floods only occur when the flow is high and the freezeup level low or moderate. Here, $Q_{\text{max}}$ is the maximum (daily mean) discharge during the breakup period (defined as the period starting with the rise of runoff and ending on the last day of “B”; the B-label indicates backwater effect due to ice according to WSC). One might consider instead $Q_b$, the flow (based on daily mean data but adjusted by interpolation for time of day) associated with the peak breakup water level at Peace Point, but $Q_{\text{max}}$ is preferred in the present context because breakup and jamming in the Delta reach of Peace River comes later than at Peace Point; therefore $Q_{\text{max}}$ better expresses the potential value of $Q_b$ for that reach. In most years, $Q_{\text{max}} > Q_b$, but there are some years where $Q_{\text{max}} < Q_b$. The latter situation would typically arise where an ice jam releases at Peace Point, but flow continues to rise while some backwater is still felt at the gauge owing to jamming farther downstream.

The data points plotted in Fig. 2 derive largely from the studies by Prowse et al. (1996) and Beltaos et al. (2006) who examined the Peace Point hydrometric records up to the years 1993 and 2002, respectively. Data for subsequent years were filled-in as part of the present study and are summarized in Table 3.

5. Use of conditional probabilities to quantify climate and regulation impacts

Before proceeding with the derivation of mathematical equations to assess how climate and regulation have each contributed to the reduced frequency of ice-jam flooding of the PAD, it is appropriate to define certain key parameters and terms:

$H_R$ = a reference value of $H_F$, such that there is a relatively high probability of an ice-jam flood occurring when $H_F < H_R$. Beltaos et al. (2006) suggested the value of 213.4 m for $H_R$, while Fig. 2 shows that 5 of the 6 floods that occurred within the period spanned by the hydrometric record at Peace Point were associated with freezeup levels that were lower than 213.4 m. The inhibiting effect of the higher $H_F$ value for the remaining flood event (1997) appears to have been offset by extremely high breakup flows.

$Q_o$ = a reference value of $Q_{\text{max}}$, such that no ice-jam floods occur for $Q_{\text{max}} < Q_o$. Beltaos (2003a) used the numerical model RIVJAM to find that 4000 m$^3$/s is a minimum flow requirement for overbank flooding. This figure actually applies to $Q_o$, so that the corresponding $Q_{\text{max}}$ may be somewhat higher. Moreover, a spring hydrograph with a peak of 4000 m$^3$/s could only result in minimal and very brief flooding. For an ecologically meaningful flood to occur, the threshold value should be exceeded for a few days; this implies that the actual peak flow may have to be well above 4000 m$^3$/s. This is also illustrated in Fig. 2, where $Q_{\text{max}}$ is over 5000 m$^3$/s for all flood events. Though not shown in Fig. 2, the lowest recorded $Q_o$ value associated with a flood event is 4000 m$^3$/s.

“Flood”: Herein, this term should be understood to imply an ice-jam flood that results in significant replenishment of the perched basins of the Peace Delta. Within the range of the hydrometric record at Peace Point, such floods have occurred in 1963, 1965, 1972, 1974, 1996, and 1997. Timoney’s (2009) compilation of historical PAD floods, which is partly based on the report by Peterson (1994), designates these events as “large ice-jam floods”. Detailed descriptions of the 1996 and 1997 flood events can be found in Giroux (1997a, 1997b).

$P(A)$: probability of event or condition $A$, occurring in any one year. In probability notation, the expression $P(A|B)$ denotes the probability of $A$ occurring, given that $B$ has occurred; the expression $P(A|B')$ denotes the probability of $A$ and $B$ occurring at the same time.

Table 3

<table>
<thead>
<tr>
<th>Breakup year</th>
<th>$H_F$ (m)</th>
<th>$Q_{\text{max}}$ (m$^3$/s)</th>
</tr>
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<tbody>
<tr>
<td>2003</td>
<td>215.22</td>
<td>5770</td>
</tr>
<tr>
<td>2004</td>
<td>214.62</td>
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<td>2005</td>
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<td>2007</td>
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<td>2008</td>
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<td>2009</td>
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<tr>
<td>2010</td>
<td>213.85</td>
<td>2130</td>
</tr>
<tr>
<td>2011</td>
<td>213.61</td>
<td>4680</td>
</tr>
<tr>
<td>2012(*)</td>
<td>214.47</td>
<td>2870</td>
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Table 2

<table>
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<th>Variable</th>
<th>2003</th>
<th>1972</th>
<th>1996</th>
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<td>$H_F$ (m)</td>
<td>215.22</td>
<td>213.01</td>
<td>212.38</td>
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<tr>
<td>$H_o$ (m)</td>
<td>1.23</td>
<td>1.04</td>
<td>0.87</td>
</tr>
<tr>
<td>$Q_{\text{max}}$ (m$^3$/s)</td>
<td>5770</td>
<td>5600</td>
<td>5800</td>
</tr>
<tr>
<td>Resistance parameter, $r_d$</td>
<td>0.041</td>
<td>0.020</td>
<td>0.015</td>
</tr>
</tbody>
</table>

(*) Last year for which published (online) flow data were available at the time of writing.
With this background, the probability of a flood can be written as:

\[ P(\text{flood}) = P(\text{flood} \cap (Q_{\text{max}} > Q_o)) + P(\text{flood} \cap (Q_{\text{max}} \leq Q_o)) \]  

(3)

By definition, no floods occur when \( Q_{\text{max}} \) does not exceed \( Q_o \). Therefore the second term on the RHS (Right-Hand-Side) of Eq. (3) is equal to zero. Expanding the first term on the RHS, Eq. (3) can be re-stated as:

\[ P(\text{flood}) = P(\text{flood} \cap (Q_{\text{max}} > Q_o) \cap (H_F < H_0)) + P(\text{flood} \cap (Q_{\text{max}} > Q_o) \cap (H_F \geq H_0)) \]

(4)

Using the well-known conditional probability expression \( P(A \cap B) = P(A)P(B) \), Eq. (4) can be re-arranged to read (after some algebra)

\[ P(\text{flood}) = P_1 \times P_2 \times P_F + P_2 \times P_Q \times (1-P_F) \]

in which

\[ P_Q = P(Q_{\text{max}} > Q_o) \]

\[ P_F = P(H_F < H_0) \]

(6)

and

\[ P_1 = P(\text{flood} \cap (Q_{\text{max}} > Q_o) \cap (H_F < H_0)) \]

\[ P_2 = P(\text{flood} \cap (Q_{\text{max}} > Q_o) \cap (H_F \geq H_0)) \]

(7)

To arrive at Eq. (5), it was assumed that \( Q_{\text{max}} \) and \( H_F \) are independent variables so that \( P(C \cap D) = P(C)P(D) \). It could be argued that there may be a possible correlation between these variables because the “ice clearing” flow is expected to increase with increasing freezeup stage \( (\text{Beltaos, 2008a}) \). However, \( Q_{\text{max}} \) is simply the highest flow that occurs during the breakup event and may not be necessarily equal to the ice clearing flow. This is illustrated in Fig. 2, where the scatter of the data points overwhelms any correlation that might be present. A linear regression for the entire set of data points indicated a negligibly small trend \((-30 \text{ m}^3/\text{s} \text{ per m of freezeup stage}) \) with \( R^2 = 0.00033 \).

By factoring out \( P_Q \), Eq. (5) can be written more compactly:

\[ P(\text{flood}) = P_1 \times (P_2 - P_F)P_F + P_2 \]

(8)

This equation clearly delineates the effects of the two variables, \( Q_{\text{max}} \) and \( H_F \). The first term quantifies the effect of discharge, while the second term quantifies the effect of freezeup level.

Writing Eq. (8) for both the pre- and post-regulation periods and dividing through, results in

\[ \frac{P(\text{flood})_P}{P(\text{flood})_N} = \frac{P_R}{P_Q} \left( \frac{P_1 - P_2}{P_1} \right) \frac{P_{FR}}{P_{FN}} + P_2 = R_Q \times R_F \]

(9)

where the subscripts \( N \) and \( R \) respectively denote natural and regulated flow conditions. The probabilities \( P_1 \) and \( P_2 \) reflect properties of the river system, as elaborated in the following section. The ratio on the LHS of Eq. (9) quantifies the change in flood frequency, so that a value less than 1 indicates drying. In the present context, the PAD has experienced drying; therefore the LHS ratio is expected to be less than 1.

Considering next the RHS of Eq. (9), one may note that the first fraction \( (R_Q) \) quantifies the contribution of the reduced spring flow, and thence of climate, to the drying of the PAD; while the second fraction \( (R_F) \) quantifies the effect of the increased freezeup level, and thence of regulation \( (\text{Beltaos et al., 2006}; \text{ see also Section 8}) \). The smaller the value of either one of these fractions, the greater is its respective contribution to the drying trend. It is assumed for the present that regulation has had no discernible impact on the breakup flows in lower Peace River, as was concluded by Beltaos et al. (2006). It is recognized, however, that earlier investigators suggested otherwise \( (\text{Prowse and Conly, 1998}; \text{ Prowse et al., 1996}) \). This aspect of the Peace River ice regime is explored in detail later \( (\text{Section 7}) \) and an additional equation is developed to account for regulation-induced changes to breakup flow.

6. Test of the probabilistic method

One can test the preceding approach by considering how well Eq. (8) predicts \( P(\text{flood}) \) in the post-regulation period, which consists of 41 years (1972–2012; the reservoir-filling years 1968–1971 are excluded as not representative of the post-regulation regime; 2012 is the last year for which Peace Point hydrometric data have been published by WSC). The probabilities \( P_1 \) and \( P_2 \) can be estimated by ranking recorded values of \( H_F \) and \( Q_{\text{max}} \) \( (\text{Figs. 3 and 4}) \) and assigning respective probabilities via a plotting-position formula. One of the simplest formulae, the Weibull equation, has been adopted herein \( (\text{Makkonen, 2006}; \text{ White and Beltaos, 2008}) \):

\[ P = \frac{r}{n+1} \]

(10)

in which \( r = \text{rank in the ascending series; } n = \text{ total number of available values; and } P = \text{probability of non-exceedance}. \)

The “observed” values of \( P(\text{flood}) \) can be determined in a similar manner. In the post-regulation period, there have been 4 flood events, as noted earlier, while the “usable” post-regulation record amounts to 41 years. Therefore, \( P(\text{flood})_P = 0.095 \left[ = \frac{4}{41 + 1} \right] \), per Eq. (10). Similar calculations for natural flow conditions yield different results, depending on how far back in time the pre-regulation record is extended. For instance, if the period of hydrometric data availability at Peace Point is used \( (1960–1967) \), there are 2 flood events in 8 years, or \( P(\text{flood})_N = 0.22 \left[ = \frac{2}{8 + 1} \right] \). If the record is extended back 41 years, same as the regulated-flow period, there are 8 flood events, and hence \( P(\text{flood})_N = 0.19 \left[ = \frac{8}{41 + 1} \right] \). In both cases, \( P(\text{flood})_P / P(\text{flood})_N < 1 \) (drying).

The more complex probabilities \( P_1 \) and \( P_2 \) can be conveniently determined from the entire data set of Fig. 1. As noted earlier \( (\text{Section 3}) \), \( P_1 \) and \( P_2 \) are considered properties of the specific river system. This is based on the fact that they already account for the effects of regulation and climate via those combinations of the main governing variables, \( Q_{\text{max}} \) and \( H_F \), that can result in flooding. In addition to freezeup level and breakup flow, ice cover thickness and climatic conditions during the period of ice breakup also influence these probabilities. However, thickness \( (\text{solid-ice sheet}) \) has changed very little since the 1960s, despite a mild warming that has been experienced in northern Alberta during the winter months \( (\text{Beltaos, 2008b}) \). At the same time, air temperatures for the month of May, which is when breakup typically occurs, have not changed during the period of the Peace Point hydrometric record \( (\text{Fig. 5}) \). On occasion, breakup may also occur during the latter part of April, but April temperatures have not changed significantly, either. Unusually cold Aprils in the 1960s \( (\text{Fig. 5}) \) would produce a slight
The value of Ho is reasonably well defined and the results are summarized in Table 4. PF and PQ are respectively equal to 0.72 and 0.48 (obtained from Figs. 6 and 7) for both natural and regulated-flow conditions. The calculated probability of ice-jam flooding, P(flood)N, works out to be 0.23, practically the same as 0.22, the value indicated by the 8-year pre-regulation record for which there are hydrometric data at Peace Point. This result provides additional corroboration of Eq. (8).

For the natural-flow period (1960–1967) and with Qo = 4300 m³/s, PF and PQ are respectively equal to 0.72 and 0.48 (obtained from Figs. 6 and 7). The calculated probability of ice-jam flooding, P(flood)N, works out to be 0.23, practically the same as 0.22, the value indicated by the 8-year pre-regulation record for which there are hydrometric data at Peace Point. This result provides additional corroboration of Eq. (8).

To calculate the various probabilities that appear on the RHS of Eq. (8), it is necessary to have the probability distributions of Hf and Qmax as well as appropriate values for Ho and Qo. The required probability distributions have been calculated using Eq. (10) and are shown plotted in Figs. 6 and 7 for both natural and regulated-flow conditions. The value of Hf is reasonably well defined at 213.4 m, as discussed earlier. For Qo, different values were tried, equal to and above 4000 m³/s, and the results are summarized in Table 4.

Best agreement between calculated and observed flood probabilities is obtained with Qo = 4300 m³/s. This result corroborates Eq. (8), as the observed flood probability is matched by the calculated value while the optimal Qo is highly plausible. With Hf = 213.4 m and Qo = 4300 m³/s, the probabilities P1 and P2, which reflect properties of the river system, regardless of regulation, are respectively calculated as 0.625 and 0.083 (cf. Table 4) using all years of record, i.e. up to 2012.

As both RQ and RF are well below 1, it is concluded that both climate and regulation have played important roles in the drying of the PAD. As the ratio RQ is higher than RF, it can also be concluded that the regulation effect on the drying of the PAD has been more pronounced than the climatic effect. Had there been no regulation, the changing climatic conditions would have caused a reduction in flood frequency equal to 21% (= (1–0.79) × 100), which accounts for about one-third of the experienced reduction of 57% (= (1–0.43) × 100). Had there been no climate-induced changes, regulation would have caused a reduction in flood frequency of 47% (= (1–0.53) × 100).

A similar calculation for the regulated-flow period ending in 2003 indicated larger values for the R-indices, suggesting that conditions have worsened in the past 10 years or so. This is consistent with the data shown in Fig. 3 with respect to regulation: the last year with HF being below 213.4 m was 2000. Fig. 4 suggests a similar, though less drastic, situation with respect to climate: Qmax has occasionally exceeded 4300 m³/s in recent years, but not often.

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The third row of Table 5 shows the values of the various parameters that result when Eq. (9) is applied to the longer-term natural flow period, 1927–1967, for which the observed flood probability is 0.19. The calculated value for the 1960–67 period is 0.23, leading to a discrepancy between the calculated and observed flood probability ratios (0.42 versus 0.50). This discrepancy can be eliminated by setting \( P_{Qo} = 0.40 \), as indicated in the last row of Table 5. The corresponding value of \( R_Q \) increases to 0.95, meaning that the reduction in flood frequency between the periods 1927–67 and 1972–2012 would be almost entirely due to regulation. The lower value of \( P_{Qo} \) relative to what is indicated by the Peace Point record (0.40 versus 0.48) is likely due to higher-than-typical breakup flows during the 1960s. Though the natural-flow record for Peace Point is too short in this regard, qualitative evidence can be gleaned by examining data at other gauges in the Peace River basin. For instance, mean April and May flow data for the earlier-established (1915) gauge at the town of Peace River (~410 km below the Bennett Dam; Table 1) indicate that natural flows were generally higher during the 1960s than before, though there is a large gap in the record between 1931 and 1958. Similar findings apply to the Smoky River, the largest tributary of the Peace (Fig. 8). Though the Peace River record also contains a gap, the flow is natural; therefore 1960s flows can be meaningfully compared to flows before and after 1968.

As noted earlier, the preceding comparisons do not account for possible effects of regulation on breakup flows at Peace Point. This is consistent with the conclusion reached by Beltaos et al. (2006), based on examination of Peace River and Smoky River flow trends as well as naturalized flow hydrographs computed by Peters and Prowse (2001). On the other hand, Prowse and Conly (1998) reported a regulation-induced increase in the flows associated with the peak breakup stage at Peace Point. This was based on flows at the Hudson Hope gauge (Table 1), located ~ 30 km downstream of, and essentially registering the outflow from, the Bennett Dam. Using the dates when peak breakup stages occurred at Peace Point and an estimated flow travel time of 7 days from Hudson Hope, Prowse and Conly (1998) found an average post-regulation increase of ~500 m³/s. This estimate relied on the six-year pre-regulation record at Peace Point for which water levels are available (1962–1967) and on post-regulation data up to 1993 (1972–1993).

At the same time, relevant observations and historical information indicate that such estimates are highly sensitive to the timing of the peak breakup stage (typically occurring in the first part of May) and the assumed 7-day travel time, which actually changes with discharge and ice conditions between the Bennett Dam and Peace Point. At Hudson Hope, the last week of April marks the transition from positive to negative regulation effects on flow, as illustrated in Fig. 9. Even a delay of a few days, as is the usual interval between breakup at Peace Point and the Delta reach of Peace River (where flood-generating ice jams occur) may mean a flow reduction, rather than increase, in the Bennett Dam contribution. Though it is difficult to discern a regulation-caused increase or decrease in breakup flows of the lower Peace River, it is of interest herein to examine how such a change might influence the comparison between the effects of climate and regulation. Letting the regulation-induced change be denoted by \( \Delta Q \), Eq. (9) is re-arranged to read:

\[
P_{\text{flood}}(\max)_{R} = R_F \frac{P(Q_{\max} + \Delta Q > Q_o)_{R}}{P(Q_{\max} > Q_o)_{R}}
\]

This equation is obtained from Eq. (9) by dividing both the numerator and the denominator of the RHS by the probability \( P(Q_{\max} + \Delta Q > Q_o)_{R} \). As noted earlier, \( R_F \) represents the effect of regulation via the increase in the freezeup level \( H_F \). The second ratio on the RHS of Eq. (11) represents the effect of regulation on ice-jam flooding via the flow change \( \Delta Q \), since \( (Q_{\max} + \Delta Q) \) expresses what the post-regulation discharge would have been if unaffected by climate; this ratio will exceed 1 (a positive effect on ice-jam flood frequency) if

Table 4
Calculated versus observed flood probabilities for post-regulation period, 1972–2012.

<table>
<thead>
<tr>
<th>( Q_o ) (m³/s)</th>
<th>( P_{FR} )</th>
<th>( P_{Qo} )</th>
<th>( P_1 )</th>
<th>( P_2 )</th>
<th>Calculated ( P ) (flood)</th>
<th>Observed ( P ) (flood)</th>
<th>Error (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>4000</td>
<td>0.31</td>
<td>0.46</td>
<td>0.455</td>
<td>0.059</td>
<td>0.083</td>
<td>0.955</td>
<td>-12</td>
</tr>
<tr>
<td>4100</td>
<td>0.31</td>
<td>0.42</td>
<td>0.455</td>
<td>0.085</td>
<td>0.089</td>
<td>0.955</td>
<td>-6</td>
</tr>
<tr>
<td>4200</td>
<td>0.31</td>
<td>0.40</td>
<td>0.455</td>
<td>0.083</td>
<td>0.100</td>
<td>0.955</td>
<td>+5</td>
</tr>
<tr>
<td>4300</td>
<td>0.31</td>
<td>0.38</td>
<td>0.455</td>
<td>0.083</td>
<td>0.095</td>
<td>0.955</td>
<td>0</td>
</tr>
<tr>
<td>4400</td>
<td>0.31</td>
<td>0.35</td>
<td>0.625</td>
<td>0.083</td>
<td>0.090</td>
<td>0.955</td>
<td>-5</td>
</tr>
<tr>
<td>4500</td>
<td>0.31</td>
<td>0.33</td>
<td>0.625</td>
<td>0.091</td>
<td>0.085</td>
<td>0.955</td>
<td>-10</td>
</tr>
<tr>
<td>4600</td>
<td>0.31</td>
<td>0.30</td>
<td>0.625</td>
<td>0.111</td>
<td>0.084</td>
<td>0.955</td>
<td>-11</td>
</tr>
<tr>
<td>4700</td>
<td>0.31</td>
<td>0.29</td>
<td>0.625</td>
<td>0.143</td>
<td>0.085</td>
<td>0.955</td>
<td>-10</td>
</tr>
</tbody>
</table>

Table 5
Ratios \( R_Q \) and \( R_F \), quantifying the respective effects of climate and regulation on the frequency of ice-jam floods, \( Q_o = 4300 \) m³/s, \( H_s = 213.4 \) m.

<table>
<thead>
<tr>
<th>Pre-regulation</th>
<th>( P_{Qo} )</th>
<th>( P_{FR} )</th>
<th>( P_{Qo} )</th>
<th>( P_{FR} )</th>
<th>( R_Q )</th>
<th>( R_F )</th>
<th>( P_{\text{flood}}(\max)_{R} )</th>
<th>Calculated</th>
<th>Observed</th>
</tr>
</thead>
<tbody>
<tr>
<td>1960–67</td>
<td>0.72</td>
<td>0.31</td>
<td>0.48</td>
<td>0.38</td>
<td>0.79</td>
<td>0.53</td>
<td>0.42</td>
<td>0.43</td>
<td></td>
</tr>
<tr>
<td>1927–67</td>
<td>0.72</td>
<td>0.31</td>
<td>0.48</td>
<td>0.38</td>
<td>0.79</td>
<td>0.53</td>
<td>0.42</td>
<td>0.50</td>
<td></td>
</tr>
<tr>
<td>1927–67</td>
<td>0.72</td>
<td>0.31</td>
<td>0.40(1)</td>
<td>0.38</td>
<td>0.95</td>
<td>0.53</td>
<td>0.50</td>
<td>0.50</td>
<td></td>
</tr>
</tbody>
</table>

(1) Value that results in a more realistic estimate of \( P_{\text{flood}}(\max)_{R} \) for the period 1927–67.
Table 6

<table>
<thead>
<tr>
<th>ΔQ</th>
<th>Q₁ − ΔQ</th>
<th>Rf</th>
<th>P₁,𝑁</th>
<th>P₁,𝑅</th>
<th>P₁,ΔQᵢ +ΔQ</th>
<th>R₁,regulated</th>
<th>R₁,climate</th>
</tr>
</thead>
<tbody>
<tr>
<td>−600</td>
<td>4900</td>
<td>0.53</td>
<td>0.48</td>
<td>0.38</td>
<td>0.43</td>
<td>0.47</td>
<td>0.88</td>
</tr>
<tr>
<td>−300</td>
<td>4600</td>
<td>0.53</td>
<td>0.48</td>
<td>0.38</td>
<td>0.48</td>
<td>0.50</td>
<td>0.84</td>
</tr>
<tr>
<td>0</td>
<td>4300</td>
<td>0.53</td>
<td>0.48</td>
<td>0.38</td>
<td>0.48</td>
<td>0.53</td>
<td>0.79</td>
</tr>
<tr>
<td>300</td>
<td>4000</td>
<td>0.53</td>
<td>0.48</td>
<td>0.38</td>
<td>0.52</td>
<td>0.57</td>
<td>0.73</td>
</tr>
<tr>
<td>600</td>
<td>3700</td>
<td>0.53</td>
<td>0.48</td>
<td>0.38</td>
<td>0.56</td>
<td>0.62</td>
<td>0.69</td>
</tr>
<tr>
<td>2200</td>
<td>2100</td>
<td>0.53</td>
<td>0.48</td>
<td>0.38</td>
<td>0.90</td>
<td>0.99</td>
<td>0.42</td>
</tr>
</tbody>
</table>

(1) In determining relevant probabilities from Fig. 6, it is simpler to work with the expression P(Q max + ΔQ > Q₁) rather than with the equivalent P(Q max > Q₁).

ΔQ > 0 and vice-versa. The third ratio quantifies the effect of climate on the regulation-changed flows ||Q max + ΔQ i||N and therefore expresses the true climatic component of the change in ice-jam flood frequency.

The composite, or net, effect of regulation is therefore given by:

\[ R_{\text{net}} = \frac{R_{\text{regulated}}}{R_{\text{climate}}} = \frac{P_{Q_{\text{max}} + \Delta Q > Q_{\text{lim}}}}{P_{Q_{\text{max}} + \Delta Q > Q_{\text{lim}}}} \]

and the effect of climate:

\[ R_{\text{climate}} = \frac{P_{Q_{\text{max}} + \Delta Q > Q_{\text{lim}}}}{P_{Q_{\text{max}} + \Delta Q > Q_{\text{lim}}}} \]

Table 6 summarizes the regulation-climate comparisons that result for different scenarios concerning ΔQ values, both positive and negative.

As expected, values of R_{\text{regulated}} increase and those of R_{\text{climate}} decrease as ΔQ increases. Within the range ΔQ = −600 to +600 m³/s, R_{\text{regulated}} remains smaller than R_{\text{climate}}, meaning that the effect of regulation predominates, even when ΔQ = 600 m³/s. The last row of Table 6 shows that it would take a flow increase of 2200 m³/s to completely offset the effect of increased freezeup levels (R_{\text{regulated}} = 0.99 ≈ 1). It can be further determined, via interpolation, that the effects of regulation and climate would be equally important for ΔQ ≈ 750 m³/s.

8. Discussion

It has been shown in the preceding sections that the relative impacts of regulation and climate can be quantified and delineated by means of a mathematical analysis that makes extensive use of conditional probabilities associated with driving and resisting forces of breakup and jamming events. The driving forces were indexed by the peak breakup discharge, Q_{\text{max}}, while the resisting forces were indexed by the freezeup level, H_{\text{f}}. Physical reasoning has indicated that the maximum winter ice thickness (h_{\text{w}}) is also relevant to the resisting forces (Eqs. (1) and (2)), though its effect is secondary to that of H_{\text{f}}. Moreover, previous work (Beltaos, 2008b) found that h_{\text{w}} has changed little in the post-regulation period. Together, these results point to the freezeup level as the dominant resistance variable that can influence ice-jam flood frequency.

Regulation is an obvious cause of the increase in H_{\text{f}} that has been experienced during the past 40+ years (Fig. 3). In principle, a changing climate could also have been a contributing factor. However, mean flow data for the late fall/early winter months (November, December, January) in two unregulated tributaries, the Smoky and Wabasca Rivers (Table 1), indicate minor, if any, climatic effects on freezeup levels (Beltaos et al., 2006). This finding is further reinforced by more recent data, extending to the year 2011, as illustrated in Fig. 10 for the largest tributary of the Peace, the Smoky River. The linear trendlines plotted for each of the two time series have very mild slopes, indicating minimal change; similar results were obtained for the month of January. The available data for the Wabasca River, a smaller tributary, extend from 1970 to 2010 and indicate slight decreasing trends for all three months of November, December, and January. This evidence suggests that climatic contributions to the increase in post-regulation values of freezeup levels are either nil or slightly negative. In the latter case, the calculated regulation effect on the frequency of ice-jam floods would be somewhat less than the true effect.

9. Summary and conclusions

Using conditional probabilities, a quantitative method has been developed to assess and compare the effects of climate and regulation on the frequency of ice-jam flood events in lower Peace River, which are known to replenish and sustain the perched basins of the PAD. The probabilistic method was tested and validated by showing that it accurately predicted observed flood frequencies during the pre- and post-regulation portions of the period for which hydrometric gauge data are available. As a by-product of the validation process, the key conditional probability P_{f} was assessed at 0.625, meaning that there is roughly a 60% chance of an ice-jam flood to occur in years when the breakup flows exceed 4300 m³/s and the freezeup level is under 213.4 m.

Assuming that regulation has had no effect on breakup flows, as concluded by Beltaos et al. (2006), it was found by means of the probabilistic method that both regulation and climate had contributed to the drying of the PAD, with regulation being the dominant factor. If the natural-flow period is limited to the years for which there are hydrometric data at Peace Point (1960–67), the present results indicate that regulation accounts for nearly two-thirds of the reduction in ice-jam flood frequency. Application of the present analysis method to the period 1927–67 (equal in length to the post-regulation record) suggested that the effect of regulation would have been much more pronounced. The latter finding is predicated on the assumption of higher 1960s breakup flows than before, which is supported by flow data at other gauges within the Peace River basin, but only in a qualitative sense.

Had regulation resulted in increased breakup flows by ΔQ = 600 m³/s (a value of 500 m³/s was suggested by Prowse and Conly, 1998), the net contribution of regulation to the drying trend would still be greater than the contribution of climate, but only slightly. In general, the calculated net contribution of regulation decreases with increasing assumed values of ΔQ and is nullified for ΔQ = 2200 m³/s.

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References


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