Geomorphology 108 (2009) 273–281

Contents lists available at ScienceDirect

Geomorphology

journal homepage: www.elsevier.com/locate/geomorph

Impacts of recurring ice jams on channel geometry and geomorphology in a small high-boreal watershed

Étienne Boucher [a,*], Yves Bégin [b], Dominique Arseneault [c]

[a] *Centre d'études nordiques and Département de géographie, Université Laval, Québec, Canada G1V 0A6*
[b] *Centre d'études nordiques and Centre Eau, Terre & Environnement, Institut national de la recherche scientifique, Québec, Canada G1K 9A9*
[c] *Centre d'études nordiques and Département de biologie, chimie et géographie, Université du Québec à Rimouski, Rimouski, Canada G5L 3A1*

ARTICLE INFO

Article history:
Received 23 October 2008
Received in revised form 2 February 2009
Accepted 3 February 2009
Available online 20 February 2009

Keywords:
Ice jams
Fluvial geomorphology
Hydraulic geometry
Dendrogeomorphology
Ice scars
High boreal

ABSTRACT

River ice jams are generally perceived as significant erosive events and are well known to impact both channel morphology and geometry. However, the extent of these impacts and the frequency of events required to maintain erosion-induced morphologies remain unexplored in most cold region watersheds. In this study, we investigated downstream variations in channel width, cross-sectional area, depth, and geomorphological characteristics in a small high-boreal basin. We coupled these observations to dendrochronological data on ice jam frequency. Our results show that channels affected by ice erosion appear enlarged and present an important retreat of the upper bank. Such enlarged channels present a typical two-level, ice-scoured morphology when ice jams recur more often than once every 5 years. By contrast, channels appear unaffected when ice jams are less frequent. These results suggest that ice jams maintain ice-scoured and enlarged morphologies once a minimal frequency-of-occurrence threshold is exceeded. We therefore conclude that ice jam frequencies should be taken into account in order to better define the role of ice as a geomorphological agent in cold environments.

1. Introduction

The size and shape of stream channels and their capacity to contain flows of various size and frequency vary according to the hydrologic regime and sediment supply (Wolman and Miller, 1960; Harvey, 1969). This useful concept forms the basis of the hydraulic geometry theory (Leopold and Maddock, 1953) from which the following principles can be drawn: (i) alluvial channels are maintained by relatively frequent flows reaching the bankfull stage, and (ii) downstream changes in channel characteristics at bankfull are related to increasing discharge or contributory area. Hydraulic geometry relations (i.e., functions relating discharge to channel characteristics) were extensively used by engineers and geomorphologists to predict channel size and to manage streams and floodplains of the temperate climate (Singh, 2003).

Whether or not these hydraulic geometry relations hold for cold region watersheds is much debated, mainly because of ice effects. From an examination of 24 rivers in Alberta (Canada), Smith (1979, 1980) proposed that repeated episodes of ice abrasion and gouging of the banks during breakup ice jams could be responsible for the maintenance of enlarged and high capacity bankfull channels. More specifically, these results suggested a region-specific discharge (Q) threshold marking the occurrence of mobile ice and ensuing ice scouring (Fig. 1A). Once this threshold is exceeded, ice abrasion on banks and bed causes a step change in geometric properties of channels such as width, cross-sectional area, or depth (Smith, 1980). More recently, McNamara (2000) and Best et al. (2005) reinforced the hypothesis of Smith (1979) by reporting a downstream hydraulic geometry anomaly in the Kuparuk watershed, Alaska, that corresponds to a transition from bedfast to floating ice (Fig. 1B).

However, in a discussion of the results of Smith (1979), Kellerhals and Church (1980) identified some potential sources of error, such as confounding fluvial terraces and genetic floodplains. These authors insisted on the fact that additional processes such as channel entrenchment, channel icings, and backwater effects behind ice jams might have engendered geometric unconformities in Albertan rivers. Moreover, the study of Smith (1979) did not include any data describing how frequent ice jams must be to maintain such enlarged channel forms. Smith (1980) suggested that, when a stream acquires the capacity to fragment and mobilize its ice cover, enlarged morphologies systematically develop. This may be an overly simplistic assumption considering the fact that ice jam frequencies vary importantly among stream reaches in a watershed (Beltaos, 1996; Boucher, 2008). We must also consider that Smith (1979) did not provide any geomorphological description, so the variety of processes leading to channel enlargement remains unclear.

---

[*] Corresponding author. Present address: Centre Eau, Terre & Environnement, Institut national de la recherche scientifique, Québec, Canada G1K 9A9. Tel.: +1 418 654 3823; fax: +1 418 654 2600.
E-mail address: etienne.boucher@ete.inrs.ca (É. Boucher).

0169-555X/$ – see front matter © 2009 Elsevier B.V. All rights reserved.
doi:10.1016/j.geomorph.2009.02.014

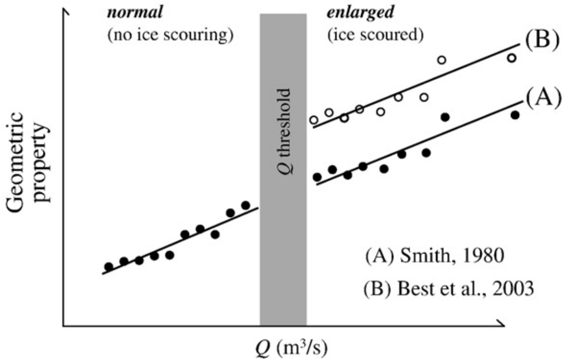

Fig. 1. Schematic representation of an idealized Q threshold separating between ice-scoured and normal channels. Adapted from (A) Smith (1980) and (B) Best et al. (2005).

Despite the important implications of the findings of Smith (1979) for fluvial geomorphology, river management, and engineering, very few additional studies intended to investigate how ice dynamics influence channel morphology and hydraulic geometry relations. An important part of the literature on river ice impacts presented geomorphological descriptions of fluvial landscapes that have apparently been shaped by ice (Dionne, 1974, 1976, 1978; Mackay et al., 1974; Hamelin, 1979; Prowse and Gridley, 1993; Dyke, 2000; Prowse, 2001; Ettema, 2002; Smith and Pearce, 2002; Walker and Hudson, 2003). However, few studies examined the spatial extent of these landforms along channels (Ettema, 2002)—a necessary aspect to assess the impact of ice dynamics on hydraulic geometry relations. Moreover, the minimal frequency of ice-scouring events to maintain such landforms remains largely unexplored, mainly because historical data on ice jam occurrence is sparse, anecdotal, and highly site-specific in most cold region watersheds. Thus, additional data is needed on bankfull geometry in rivers that undergo significant ice effects during breakup in order to verify if hydraulic geometry relations hold in these environments.

In this study, we investigated downstream variations of channel geometry and geomorphology in a small, high-boreal watershed that experiences frequent spring ice jams. We used tree-ring chronologies of ice jams, cross-sectional analysis, aerial photographs, and detailed geomorphological descriptions to determine if geometric and geomorphologic properties of channels relate to variations in the frequency of ice-scouring events.

2. Study area

The Necopastic River (53°73′ N., 78°28′ W.) is an ungauged, high-boreal watercourse located 40 km west of Radisson, James Bay, northern Québec. Its small hydroclimatically homogeneous watershed drains an area of 250 km^2, and the bankfull discharge in the lower reach is about 25 m^3/s (estimated from a rating curve). Climate normals (1971–2001) at the La Grande Rivière A station displays a typical high-boreal continental climate with mean annual temperatures around −3.1 °C (Environment Canada, 2008). Minimum (−23.2 °C) and maximum (13.2 °C) monthly averages are reached in January and July, respectively. The total annual precipitation is 684 mm, with 37% (248 mm) falling as snow. The months of December, January, and February each have an average of less than

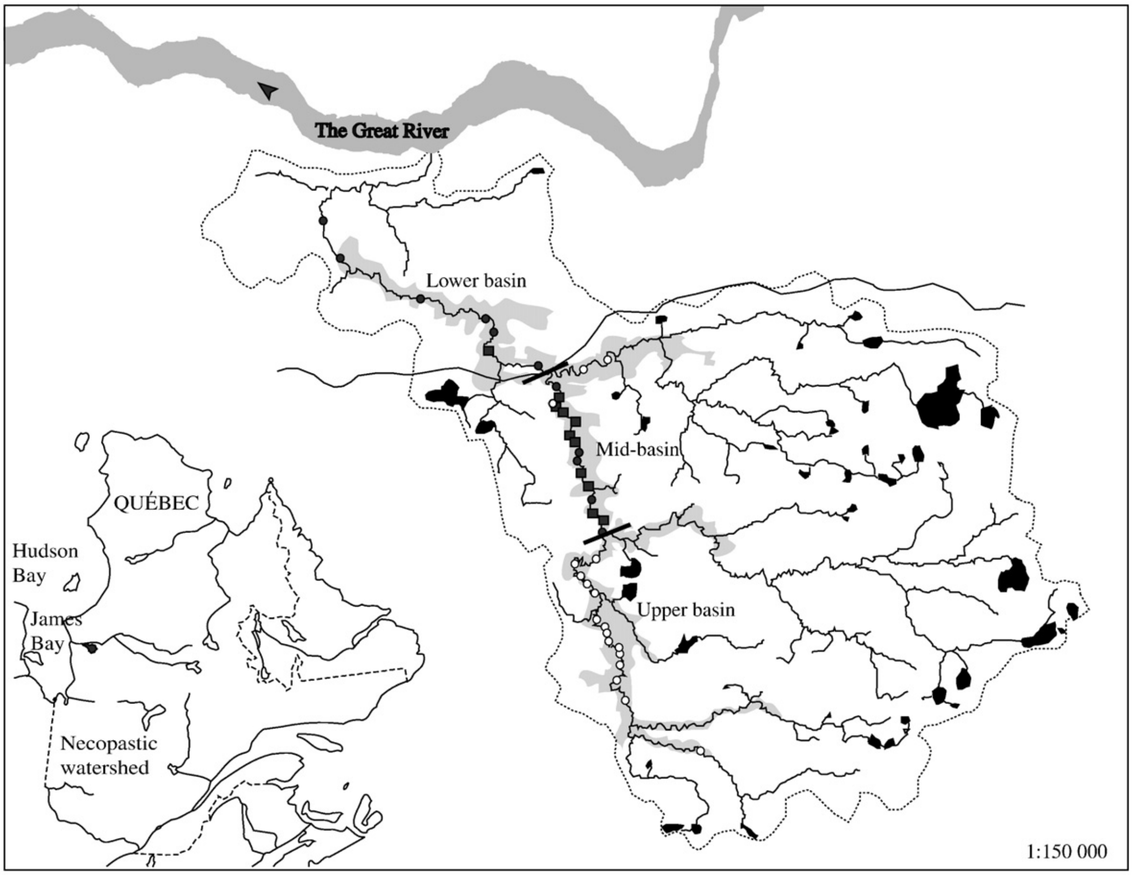

Fig. 2. The Necopastic watershed. Studied sites are represented by circles and squares. Black symbols denote sites where ice-scouring events have occurred in the past (see Fig. 3). White circles represent wide floodplain channels with no ice-scouring evidence (cluster 1, see Results). Black circles correspond to entrenched channels that experienced frequent ice jams (cluster 2). Black squares refer to entrenched sites that display an ice-scoured morphology (cluster 3). The shaded gray zone corresponds to the extent of lithologically homogeneous fluvioglacial deposits. These Holocene deposits are mainly composed of sands and silts. The symbology described here applies to the rest of the paper.

1 day with a temperature over 0 °C. In contrast with rivers of the cold-temperate climate where extreme ice jam events frequently occur during winter warm spells (Beltaos and Prowse, 2001; Beltaos and Burrell, 2003), ice jamming is most likely to occur during the spring flood on the Necopastic River.

Regional vegetation is characterized by the codominance of black spruce [*Picea mariana* (Mill. BSP)] and Jack pine [*Pinus banksiana* (Lamb)] (Arseneault and Sirois, 2004). In riparian areas, black spruce shares its dominance with eastern larch [*Larix laricina* (Du Roi) K. Koch] and balsam fir [*Abies balsamea* (L.) Mill] (Arseneault and Sirois, 2004; Bouchon and Arseneault, 2004; Boucher et al., 2006; Arseneault et al., 2007). Ice scouring tolerant shrubs such as *Alnus viridis* ssp. *crispa* (Ait.) Turrill. *Salix planifolia* Pursh. and *Betula glandulosa* Michx cover the riverbanks and floodplains of most streams in the area.

3. Materials and methods

3.1. Studied sites

The Necopastic River watershed can be separated into three classes following major increases in drainage area (upper basin, mid-basin, lower basin) and into two groups according to the type of ice cover breakup (i.e., thermal versus mechanical breakups) (Fig. 2). At each site, ice scars on shrubs (Fig. 3) indicated the presence of ice jams. A total of 39 sites were chosen along the Necopastic River: 16 in the upper basin, 16 in the mid-basin, and 7 in the lower basin. All sites are incised into geologically homogeneous Holocene fluvioglacial deposits (Vincent, 1985) (Fig. 2) and correspond to relatively rectilinear stream sections of constant slope, width, depth and cross sectional area. Sites recently flooded by the Canadian beaver (*Castor canadensis* Kuhl.) or located immediately downstream or upstream of a sill were avoided.

3.2. Geometric and geomorphologic descriptions of sites

During fieldwork, we first identified the bankfull stage that we defined as the height of the relatively flat depositional surface adjacent to the river (Wolman and Leopold, 1957; Williams, 1978). On the Necopastic River, these depositional surfaces (in some cases) correspond to the height of the valley flat (Fig. 4A). However, with entrenched channels, we carefully distinguished the floodplains from the less frequently flooded terraces. In such sites, genetic floodplains are narrow and form a lower bench that is traceable along the river channel (Fig. 4B).

Fig. 3. Shrubs scarred by drift ice.

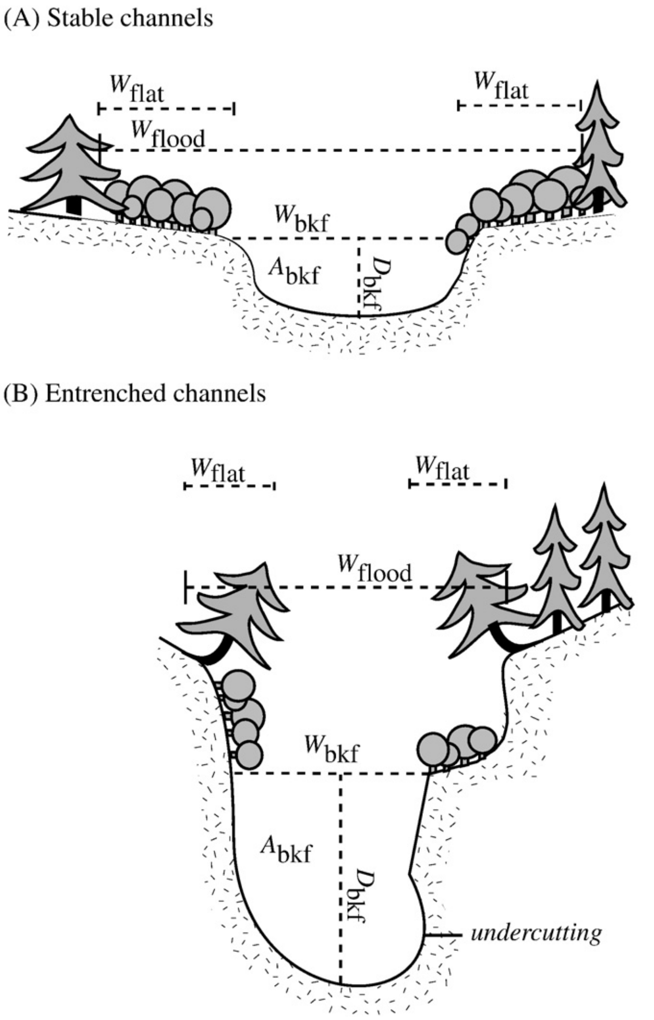

Fig. 4. Geometric properties and identification of the bankfull stage. (A) Stable and (B) entrenched channels. W_{bkf}, W_{flat}, A_{bkf}, and D_{bkf} refer to bankfull width, valley flat width, bankfull cross-sectional area, and bankfull depth, respectively.

At each site, one representative cross section was mapped using a Leica TC-705 theodolite (precision = 2 mm ± 2 ppm). These cross sections were orientated perpendicularly to streamflow direction and encompassed the full valley width. We then computed an algorithm in MATLAB R14© to interpolate between measurements and calculate bankfull channel widths (W_{bkf}), cross sectional areas (A_{bkf}), depths (D_{bkf}), and flat widths (W_{flat}) (Fig. 4). An entrenchment ratio ($ER = W_{flood}/W_{bkf}$ where W_{flood} = valley width at $2 \times D_{bkf}$) was calculated after Rosgen (1996). Theodolite measurements were finally used to evaluate the slope of the water surface and the bankfull stage. The slope values in each site were obtained by regressing a minimum of five water elevation measurements on the distance downstream. The slope gradient was very low (<0.001 m/m) in most sites (Fig. 5).

In addition to the field data, supplementary valley-flat width measurements were performed on each Strahler orders using 1:10,000 aerial photos (Société d'Énergie de la Baie James, 1984). Although multiple lithologies are encountered throughout the basin and might account for some variations, our intention was to get a general portrait of how flood-prone widths (i.e., valley flats) vary downstream. Flat widths correspond, on aerial photography, to the easily recognizable zone colonized by *Alnus viridis*, *Salix planifolia*, and *Betula glandulosa* (Bouchon and Arseneault, 2004; Boucher et al., 2006). Three width measurements were averaged in each stream section of the basin. Overall, 78 averaged widths were calculated

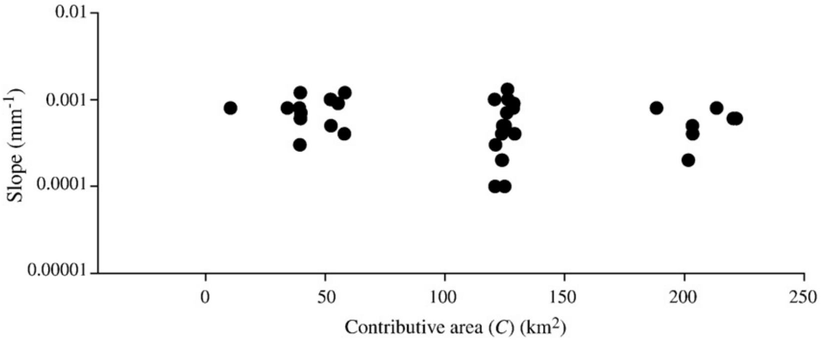

Fig. 5. Water slope values as a function of contributive area. Each dot represents a study site.

in first—($N=57$), second—($N=15$), third—($N=3$), fourth—($N=2$), and fifth—($N=1$) order streams.

All geometric measurements (W_{bkf}, W_{flat}, A_{bkf}, D_{bkf}) were plotted against the contributive area (C) instead of discharge (Q) because the Necopastic River is ungauged. Contributive area measurements were performed within PHYSITEL (Royer et al., 2006). The PHYSITEL software determines the drainage structure of a basin from a Digital Elevation Model (DEM). A DEM (cell = 20 × 20 m) of the Necopastic River watershed was constructed from SRTM elevation data (http://srtm.usgs.gov/). Hydraulic geometry power functions and equations were computed in SPSS© for Windows.

Geomorphological features and processes were described at each site using the reconnaissance survey technique (Thorne et al., 1996; Thorne, 1998). Data collected include valley description (valley height, shape, angle, coupling of channel to floodplain, valley floor type, surficial geology, number of terraces, extent of flood deposits, channel platform, lateral activity, etc.), channel description (dimensions, flow type, bed sediments, bed armouring, width controls), and left and right bank description (bank height, material, slope, shape, vegetation, status and processes of erosion and accumulation, etc.). Geometric data and geomorphologic descriptions were incorporated into a two-step cluster analysis (TSCA) to reveal groupings among sites, following a log-likelihood distance measure. Geometric data was detrended using a nonlinear regression model. Only residuals were considered in order to remove any downstream change attributable to drainage area. Furthermore, only significantly discriminating geometric or geomorphic variables ($p<0.05$ for t or χ^2 tests) will be presented and discussed here.

3.3. Tree-ring reconstructed ice jam frequencies

Trees bordering the Necopastic River present multiple ice scars (Boucher, 2008), especially at mid-basin. These scars form on trees when drift ice raised by jams comes in contact with riparian trees. These marks can be dendrochronologically dated and exhaustive event chronologies can be constructed (Payette, 1980; Smith and Reynolds, 1983; Hupp, 1988; Boucher, 2008). On the Necopastic River, scars were sampled in 15 out of the 16 mid-basin sites. None were sampled in the lower basin because this section is part of a protected area. The number of trees sampled varied considerably between sites (Table 1). To overcome this problem, Boucher (2008) established an iterative sampling-with-replacement (ISR) procedure to determine if the at-a-site sampling effort was sufficient. Three sites presented an insufficient sampling and were not considered here (Table 1).

In the field, when ice marks were well defined, cross sections were taken in the middle of each scar. Otherwise, when multiple scars were present on the trunk, cross sections were taken at multiple heights in order to date every possible scar on the tree. Trees with a diameter <10 cm were never sampled. Individuals were sampled on both banks and were separated by a minimal distance of one tree height to avoid redundancy in the record. In the laboratory, cross sections were finely sanded and tree rings were counted from the last year of growth (2005 or 2006) to the center. Dead samples were cross-dated with master chronologies existing in the area. Ice-scouring events were dated with the precaution of not replicating events found at multiple heights on a same individual. Scars that were formed after year A.D. 2003 were excluded because newly formed damages are difficult to observe in the field.

For each site, ice jam frequency was calculated by dividing the number of events recorded by the period available for recording. Sites were considered available for recording when at least one tree had grown over 3 cm of radius. We attributed a frequency of “zero” to one mid-basin site that did not show any geomorphological nor tree-ring evidence of ice scouring (Fig. 2).

4. Results

In the Necopastic River watershed, an important part (>50%) of the variations in W_{bkf} and A_{bkf} is explained by the downstream increase in contributive area (Fig. 6A,B). However, only a weak relationship is observed with channel depth (D_{bkf}) (Fig. 6C). Furthermore, from a detailed examination of power functions modeling downstream changes in width and cross-sectional area (Fig. 6A,B), considerable variability is observed at mid-basin (for contributive areas of 100 to 150 km²). For example, in the upper basin, average W_{bkf} and A_{bkf} are 7.4 ± 1.6 m and 7.9 ± 3.6 m², respectively. At mid-basin, these values respectively rise to 19.7 ± 4.8 m and 24.2 ± 8.7 m², therefore approaching values of the lower basin (i.e., 19.3 ± 3.5 m and 23.4 ± 5.7 m²). Observed downstream channel geometry variations are not related to changes in channel gradient ($r_s<0.2$, $p>0.05$ with W_{bkf}, A_{bkf}, and D_{bkf}).

Mid-basin variations in W_{bkf} and A_{bkf} are significantly related to ice jam frequencies ($r_s=0.61$ and 0.54, respectively; Fig. 7A,B), while

Table 1
Characteristics of study sites where dendrochronological sampling was conducted; missing sites were eliminated because of insufficient sampling (Boucher, 2008).

Site	Date of site availability (year A.D.)	Number of trees sampled (N)	Frequency (N y^{-1})
1	1869	18	0.66
2	1945	15	0.41
3	1871	11	0.14
4	1936	20	0.24
5	1914	29	0.51
6	1798	9	0.23
7	1856	7[a]	0.20
10	1865	8	0.25
11	1917	5[a]	0.13
12	1877	15	0.23
13	1870	14	0.18
14	1879	8	0.23
16	–	–	0
Mean			0.26

[a] All trees bearing scars have been exhaustively sampled in these sites.

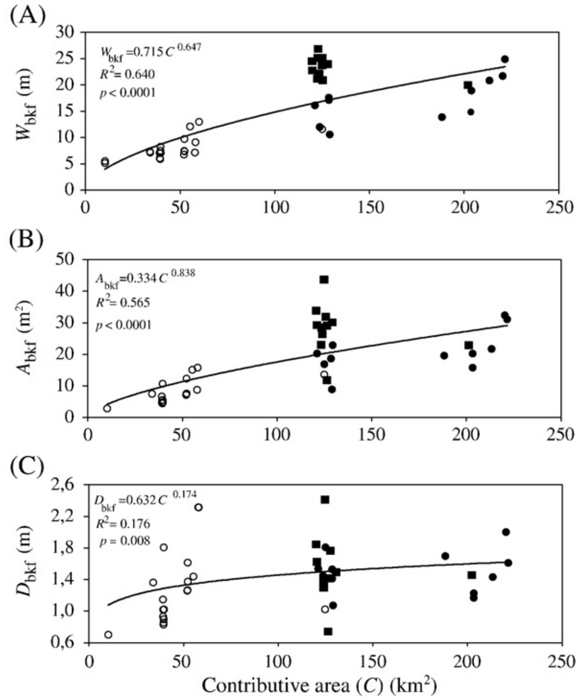

Fig. 6. Downstream changes in geometric properties of channels (W_{bkf}, A_{bkf}, D_{bkf}). Symbols correspond to those used in Fig. 2.

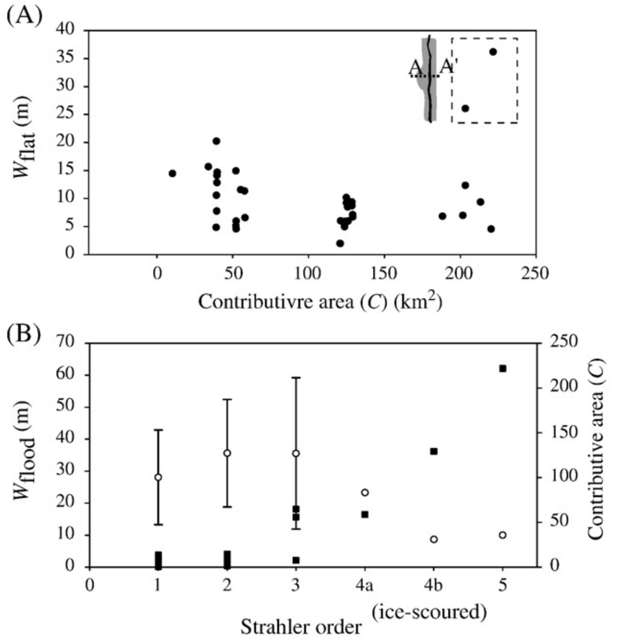

Fig. 8. Downstream changes in valley flat widths in the Necopastic watershed. (A) Valley flat widths measured in the studies sites. (B) Mean (± S.D.) valley flat widths measured on 1:10,000 aerial photographs for each Strahler order. Black squares represent the contributive area of each Strahler order. Order 4 was divided into 4a and 4b because ice scouring occurs in the latter but not in the former.

variations in channel depth are not. Narrow channels are associated with low ice jam frequencies (~0–0.2 event year^{-1} on the abscises) in comparison with larger channels (~0.2–0.6 event year^{-1} on the abscises).

W_{flat} tend to decrease in the downstream direction (Fig. 8A). This contrasts with the previously described tendency of bankfull channels to become larger at mid-basin (Fig. 6A,B). Upstream sites present relatively wide valley flats (11 ± 4.7 m) in comparison with narrower flood corridors of the mid (6.7 ± 2.5 m) and lower (8 ± 2.9 m) basins. The latter situation is also visible on 1:10,000 aerial photographies. W_{flat} are the narrowest in orders 4b and 5 where frequent ice jams occur (Fig. 8B).

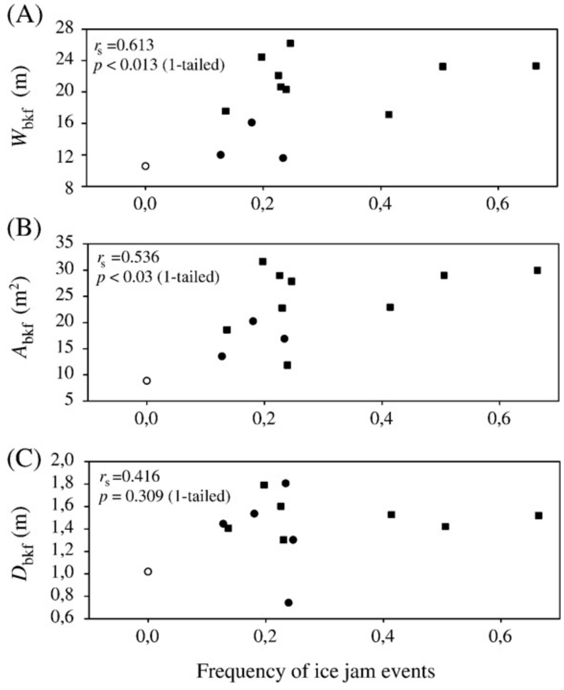

Fig. 7. Relationship between channel geometry at mid-basin and ice jam frequencies. Only sites where the sampling was sufficient were considered (see Table 1). Symbols correspond to those used in Fig. 2.

Following the TSCA, studied sites were classified into three groups according to the most influential morphological and geometrical variables. The first group (cluster 1) is found almost exclusively in the upper basin (Fig. 2) and represents wide floodplain channels (Fig. 9A). Field observations confirm the existence of extensive depositional areas adjacent to the channel in this group (Fig. 10A,B). High ER values (>2.4) point out that these sites are vertically stable (Fig. 9B). Slightly-below-zero residual widths and cross-sectional areas (Fig. 9C) indicate that the hydraulic geometry curve fits observed values quite well. Thus, we can deduce that stream sections of clusters 1 and 2 are probably laterally stable. Moreover, ice scouring activity is not present in these channels, as evidenced by the absence of ice scars on the vegetation (Fig. 9D) and no obvious ice-scoured morphology (Fig. 9E).

The two other groups (clusters 2 and 3) represent entrenched channels of the mid and lower basins. In both groups, low W_{flat} (<10 m) values and small ER ratios (<2) are indicative of incised channels (Fig. 9A,B). The omnipresence of ice scars reveals that ice jams occurred in all sites (Fig. 9D). However, a major difference between clusters 2 and 3 is that, whilst the former shows no particular evidence of lateral erosion (Fig. 10C,D), the latter depicts abnormally large channels (Fig. 9C) and presents a distinct ice-scoured morphology (Fig. 9E). Ice-scoured channels are characterised by a discontinuous and partly eroded genetic floodplain forming a narrow bench (Fig. 10E,$F_{(i)}$). This bench is covered with heavily scarred shrubs installed on freshly deposited alluvium. A steep eroded talus (Fig. 10$F_{(ii)}$) (about 50 ± 14 cm high) separates the flat depositional surface from a higher terrace. Mechanical erosion on this talus has excavated tree and shrub roots living on the higher terrace (Fig. 10$F_{(iii)}$). Recent alluviums are often found at the base of scarred trees, indicating that these terraces might slowly aggrade.

Finally, the ice-scoured morphology (cluster 3) is most often associated with sites that experienced frequent ice jams in the past. Where ice jam events occurred at least once every 5 years (frequency

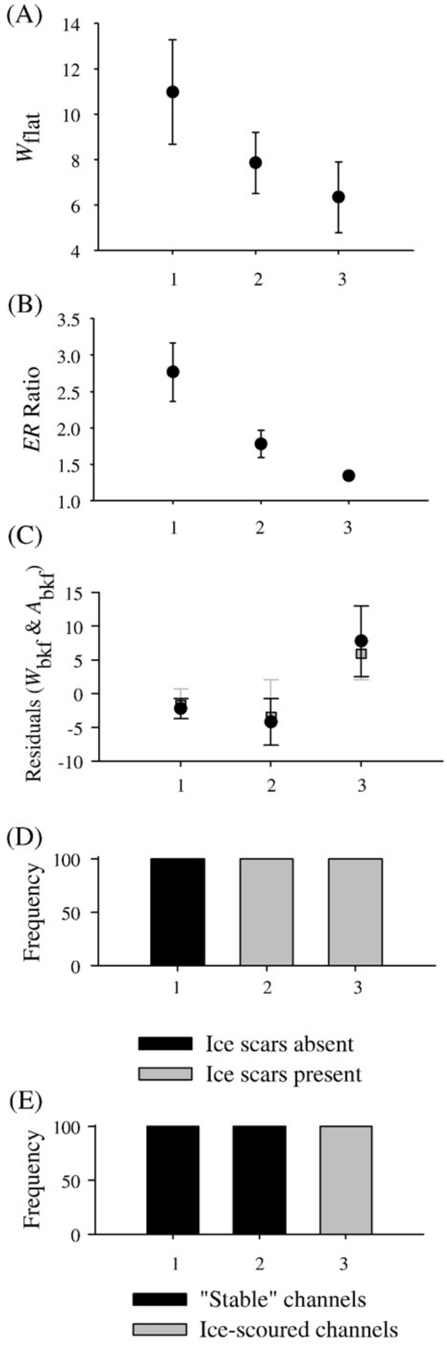

Fig. 9. Geometric and geomorphologic variables discriminating between the three clusters (1, 2, 3) produced by the TSCA. Only variables that contribute significantly to discriminate between clusters are shown [$p<0.05$ for t-tests (A, B, C) and χ^2 tests (D, E)]. Cluster 1 represents wide floodplain channels. Cluster 2 corresponds to entrenched/narrow-floodplain channels. Cluster 3 includes entrenched/ice-scoured channels.

~0.2 event year^{-1}), most sites developed an ice-scoured cross section (Fig. 7A,B).

5. Discussion

5.1. Are ice jams really causing enlargement?

In the Necopastic River watershed, recurring ice jams seem to influence the geometric and geomorphologic properties of channels, resulting in anomalously large and heavily scoured sites. However, as pointed out by Best et al. (2005), further discussion is needed to determine whether channel enlargement is truly caused by ice jams or, on the contrary, if ice movements resulting in jams are just permitted in channels that are already enlarged by other environmental processes. Among the factors other than ice that could possibly influence downstream variations in channel width are longitudinal changes in slope, lithology, and climate (Singh, 2003). In this study, the sampling strategy was specifically designed to minimize the influence of these variables on channel geometry. First, the Necopastic River basin can be considered hydroclimatically homogeneous because of its small size (<250 km^2). Second, sampling sites are all located within a Holocene fluvioglacial terrace complex composed of sandy silts (Fig. 2). As with other small streams flowing in these geologically uniform environments, the Necopastic is just reworking “previously-sorted” Holocene deposits; consequently, floodplain (modern) and terrace (inherited) sediments are indistinct granulometrically (Boucher et al., 2006). Third, downstream variations in channel slope are not related to geometric properties of channel in our system, and the gradient of all studied sites is very low (<than 0.001 mm^{-1}) (Fig. 4). Hence, downstream channel geometry variations are more likely to be attributable to characteristics of the hydrological regime (including ice effects) than to “permanent” variables affecting geomorphological processes in the watershed.

5.2. Ice jam frequency (F) thresholds

This study only partly supports the hypothesis of Smith (1979) regarding the role of ice jams as significant erosive processes in cold-environment rivers. We agree with Smith (1979, 1980) that ice jams may alter channel geometry and geomorphology, especially when channels are entrenched. However, in the Necopastic River, thresholds in the frequency (F) of ice jams seem to have a greater influence on channel geomorphology compared to discharge (Q) or contributive area (C) thresholds documented in earlier works (Smith, 1979, 1980; McNamara, 2000; Best et al., 2005). Such thresholds imply that, once a critical value is exceeded, streams systematically display an enlarged or scoured morphology (Fig. 1A,B) regardless of the intersite variations in ice jam frequency. In the Necopastic watershed, mechanically induced ice jams become common when the drainage area exceeds 120 km^2 (Fig. 2). However, downstream of this C threshold, all sites do not appear systematically scoured or enlarged. On the contrary, many mid-basin and almost all lower basin sites appear uninfluenced by ice (i.e., cluster 2) (Figs. 6A,B and 10C,D) despite the abundant tree-ring evidence for ice jamming in these environments. As demonstrated earlier, the enlarged morphology (cluster 3) becomes omnipresent in sites where ice jams recur more often than once every 5 years (Fig. 7A,B).

An important implication of these findings is that hydraulic geometry relations might only hold in sites where ice jams are less frequent than the F threshold. In other words, downstream variations in ice jam frequency hinder the precise estimation of geometric properties from hydraulic geometry curves in ice jam-prone basins. For example, in the Necopastic River watershed, the downstream increase in contributive area does not explain the large geometric variability occurring at mid-basin (Fig. 6A,B). Furthermore, downstream changes in geometric properties would probably be more accurately described from hydraulic geometry curves if ice-scoured channels (cluster 3) were not considered (Fig. 6A,B). To verify this, R^2 values for these models were recomputed after removing cluster 3 sites. As expected, the explained variance jumped to 0.838 and 0.704 for W_{bkf} and A_{bkf}, respectively.

Finally, channel entrenchment and basin hydrography are two factors that explain why the ice jam F threshold is exceeded only in some mid-basin sites and not downstream. Ice jams are well known to occur more frequently where channel morphology precludes ice movement, a very common situation in medium-sized flat reaches (Beltaos, 1996). At mid-basin, ice movements are constrained in a

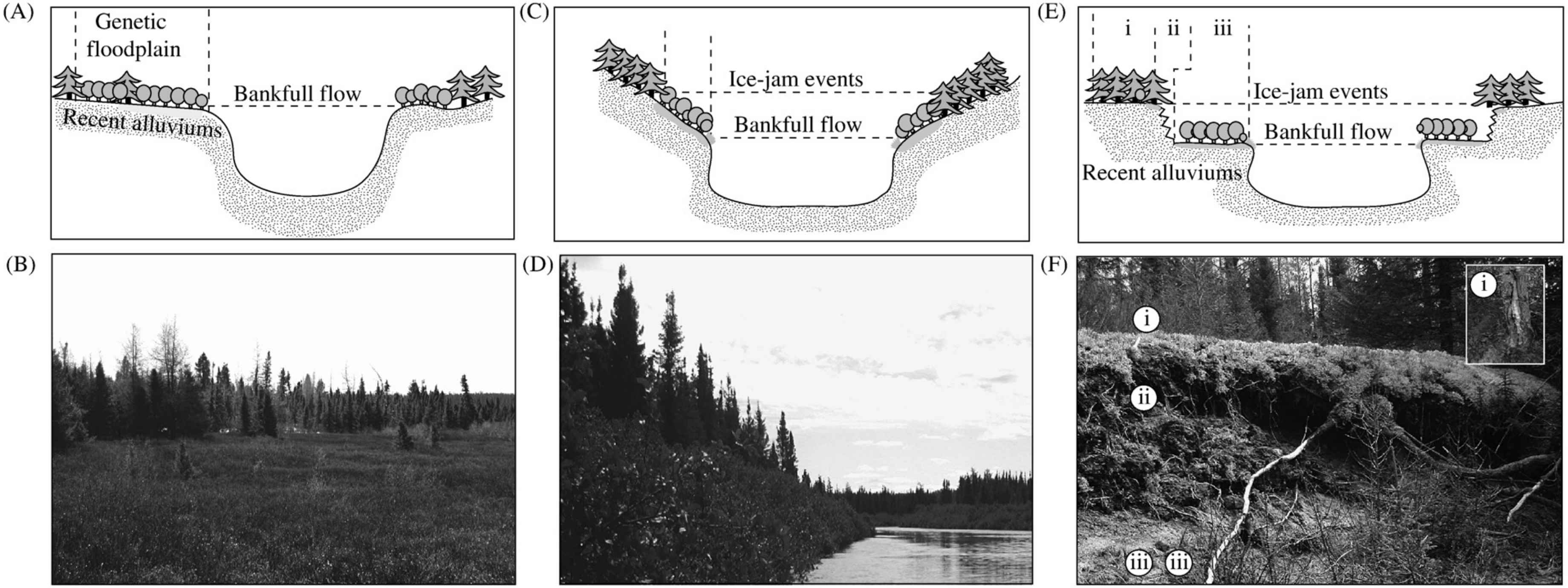

Fig. 10. Graphical and pictorial representations of the three most common geomorphological environments encountered in the Necopastic River: (A, B) wide floodplain channels; (C, D) entrenched/narrow floodplain channels, and (E, F) entrenched/ice-scoured channels. Genetic floodplains (benches), erosion talus, and ice-flood terraces with ice-scarred trees are identified as i, ii, and iii, respectively.

highly entrenched channel. Consequently, ice rafts frequently accumulate in nearby sections during mechanical breakups. In the lower basin, ice rafts possibly move farther downstream because of increased channel sizes and higher discharge. Thus, jam-prone sections tend to distance from one another so that *F* thresholds are probably rarely crossed in the lower basin.

5.3. Bank scouring processes

In sites where the *F* threshold is crossed, repeated ice-scouring events caused the riverbanks to retreat (Fig. 12A,B). Scouring in these entrenched channels is probably very severe because the system's energy cannot dissipate laterally. Moreover, the riverbank's top is extremely vulnerable to erosion as it is often free of snow and partly unfrozen at the moment of breakup (Fig. 11A). For its part, the lower bench appears constricted and very narrow but has not been completely eroded. A first explanation is that water levels rise much higher than the bankfull stage during ice jams. From a detailed analysis of ice scar heights in this watershed, it was shown that water levels rise, in average, 1 m above the bankfull stage during these events (Boucher, 2008). Remnants of the initial ice cover may also be attached to the riverbank toe, therefore conferring additional protection to the lower bench (Smith, 1980; Ettema, 2002). Nevertheless, the resulting two-level channel is similar to the one documented by Smith (1980).

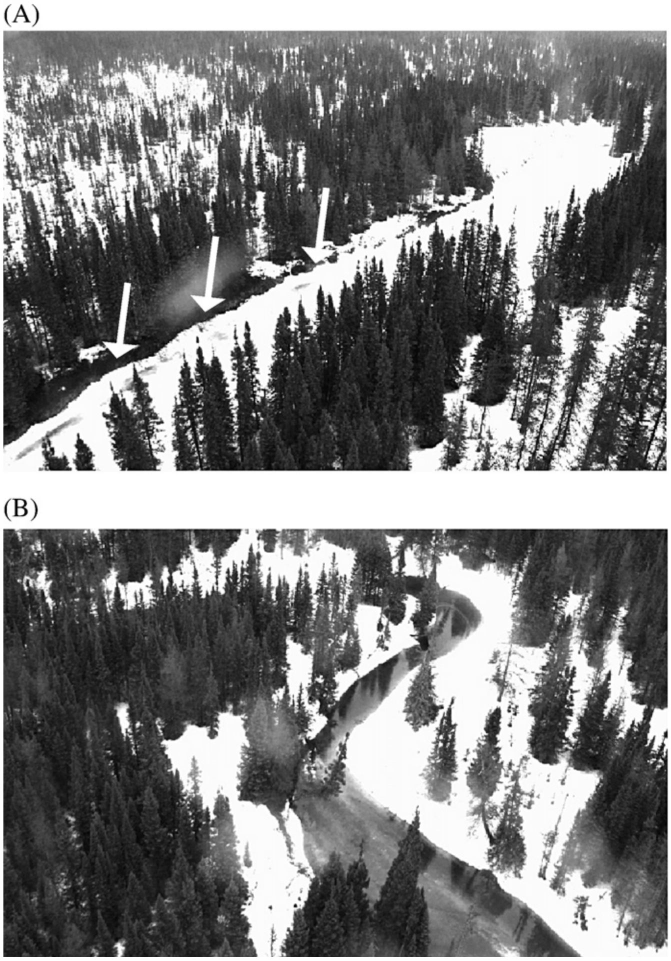

Fig. 11. The Necopastic River just before spring breakup in 2006. (A) In downstream sites, the top of riverbanks is often free of snow (white arrows), therefore augmenting their vulnerability to erosion. (B) In upstream sites, flood waters are forced on top of the ice cover and inundate the full flood-prone width without eroding the banks.

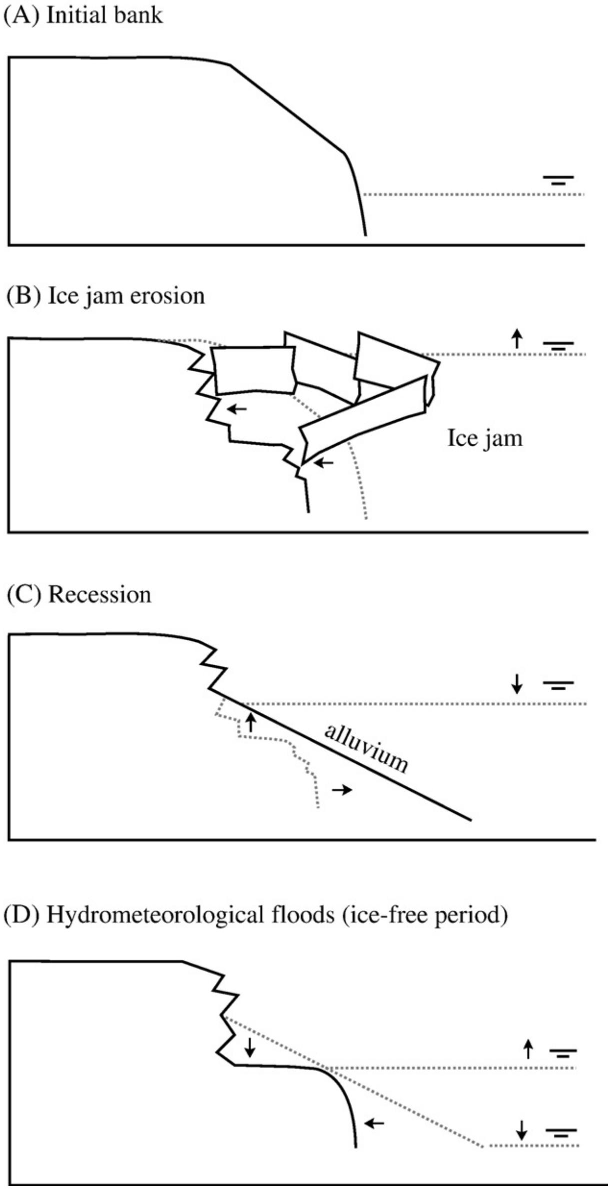

Fig. 12. Bank retreat mechanisms in portions of the Necopastic River that are frequently affected by ice jams. (A) Initial stage, (B) ice jam erosion, (C) deposition of sediments on the lower bench during flood recession, and (D) reworking of fresh alluviums by hydrometeorological floods occurring during the ice-free period in these entrenched channels.

During flood recession, sediments are deposited on the lower bench forming an inclined surface of loose material (Figs. 10F$_{(iii)}$ and 12C). These recently deposited sediments are very sensitive to erosion and are rapidly reworked into a lower bench by subsequent floods during the ice-free period (Fig. 12C). The surface of the lower bench roughly corresponds to the bankfull stage. Although vegetation encroaches and stabilizes this aggradational surface, fluvial erosion steepens and eventually undercuts the lower bench. Between each ice-scouring event, benches can experience several depositional episodes during open-channel floods.

Finally, ice dynamics may explain the presence of wide floodplains in the upper basin. In high-boreal regions, severe winters generate thick ice covers that often remain still during breakup in low-order streams. An important consequence is that spring flood waters frequently overtop the ice cover in the upper basin and inundate the full flood-prone area (Fig. 11B). The ice cover then melts in situ from thermal exchanges with flood waters and the atmosphere. In combination with flood events occurring during the ice-free period, this phenomenon probably contributes to widening the flood-prone area. A similar process was documented for the Kuparuk basin (Best et al., 2005) and is possibly representative of many small streams of the cold areas (Kellerhals and Church, 1980).

6. Conclusion

Our results demonstrate that hydraulic geometry relations lead to imprecise width (W_{bkf}) and cross-sectional area (A_{bkf}) estimations in the Necopastic River, a small stream that experiences frequent ice jams. However, the hypothesis of Smith (1979) regarding the role of ice jams as important and generalized erosive events in high-latitude rivers is only partly supported by our data. In our study, ice jams have to occur at least once every 5 years to maintain an enlarged and ice-scoured morphology. Thus, our data suggests that some frequency-of-occurrence thresholds might have to be crossed for ice jams to become geomorphologically significant in the Necopastic River basin. Indeed, a frequency–magnitude approach would be more useful than a presence/absence assessment to quantify the geomorphic impacts of these extreme events on riparian environments.

Among the visible geomorphic impacts on the Necopastic River is a “two-level” channel occurring in frequently ice-scoured sites. Our study provided new insights on the mechanisms forming these enlarged two-level channels. It is suggested that both ice jams and the yearly open-channel flood might be involved in shaping these landforms. First, ice-jam-related flooding laterally erodes the riverbank. Second, alluviums are deposited on the lower part of the riverbank during flood recession. Third, sediments are reworked into a flat bench by hydrometeorological floods occurring during the ice-free period.

Acknowledgements

We would like to thank Marilou Girard-Thomas, Mathieu Roy, and Marc-André Robin for the field assistance. This research could not have been successful without the technical assistance of Sylvain Jutras, Patrick Gagnon and Alain Rousseau. We gratefully acknowledge the following organisations for their financial support: NSERC (National Science and Engineering Research Council of Canada), FQRNT (Fonds Québécois de Recherche sur la Nature et les Technologies); Hydro-Québec, Ouranos, ArcticNet and the Northern Student Training Program of the Department of Indian and Northern Affairs of Canada. The town of Radisson (James Bay) also provided lodging facilities for this study. The quality of the original manuscript was improved by the valuable comments of two anonymous reviewers.

References

Arseneault, D., Sirois, L., 2004. The millenial dynamics of a boreal forest stand from buried trees. Journal of Ecology 92, 490–504.

Arseneault, D., Boucher, É., Bouchon, É., 2007. Asynchronous forest-stream coupling in a fire-prone boreal landscape: insights from woody debris. Journal of Ecology 95, 789–801.

Beltaos, S., 1996. River Ice Jams. Water Resources Publications, Highland Ranch, Colorado, USA. 390 pp.

Beltaos, S., Prowse, T.D., 2001. Climate impacts on extreme ice jam events in Canadian rivers. Hydrological Sciences Journal 46, 157–182.

Beltaos, S., Burrell, B.C., 2003. Climate change and river ice breakup. Canadian Journal of Civil Engineering 30, 145–155.

Best, H., McNamara, J.P., Liberty, L., 2005. Association of ice and river channel morphology determined using ground-penetrating radar in the Kuparuk River, Alaska. Arctic, Antarctic and Alpine Research 37, 157–162.

Boucher, É., 2008. Analyse hydro-climatique et géomorphologique des déglacements mécaniques de la rivière Necopastic au Québec nordique. Department of Geography, Laval University, Québec, Canada. 153 pp.

Boucher, É., Arseneault, D., Hétu, B., 2006. Late Holocene development of a floodplain along a small meandering stream, northern Quebec, Canada. Geomorphology 80, 267–281.

Bouchon, E., Arseneault, D., 2004. Fire disturbance during climate change: failure of postfire forest recovery on a boreal floodplain. Canadian Journal of Forest Research 34, 2294–2305.

Dionne, J.-C., 1974. How drift-ice shapes the St-Lawrence. Canadian Geographic Journal 88, 4–9.

Dionne, J.-C., 1976. Le glaciel de la région de La Grande rivière, Québec subarctique. Revue de Géographie de Montréal 30, 133–153 (in French).

Dionne, J.-C., 1978. Le glaciel en Jamésie et en Hudsonie, Québec subarctique. Revue de Géographie physique et Quaternaire 32, 3–70 (in French).

Dyke, L.D., 2000. Shoreline permafrost along the Mackenzie River. In: Dyke, L.D., Brooks, G.R. (Eds.), The Physical Environment of the Mackenzie Valley, North-West Territories; A Base Line for the Assessment of Environmental Change. Geological Survey of Canada, Ottawa, Canada, pp. 187–195.

Environment Canada, 2008. Canadian Climate Normals and Averages, 1971–2000. http://www.climat.meteo.ec.gc.ca/climate_normals/index_e.html?.

Ettema, R., 2002. Review of alluvial-channel responses to river ice. Journal of Cold Regions Engineering 16, 191–217.

Hamelin, L.E., 1979. The bechevnik: a river bank feature from Siberia. The Musk Ox 25, 70–72.

Harvey, A.M., 1969. Channel capacity and the adjustment of streams to hydrologic regime. Journal of Hydrology 8, 82–98.

Hupp, C.R., 1988. Plant ecological aspects of flood geomorphology and paleoflood history. In: Baker, V.R., Kochel, R.C., Patton, P.C. (Eds.), Flood Geomorphology. John Wiley & Sons, New-York, pp. 335–356.

Kellerhals, R., Church, M., 1980. Comments on “Effect of channel enlargement by river processes on bankfull discharge in Alberta, Canada” by D.G. Smith. Water Resources Research 16, 1131–1134.

Leopold, L.B., Maddock, T., 1953. The hydraulic geometry of stream channels and some physiographic implications. U.S. Geological Survey Professional Paper No. 252, Washington, DC, pp. 1–56.

Mackay, D.K., Sherstone, D.A., Arnold, K.C., 1974. Channel Ice Effects and Surface Water Velocities from Aerial Photography of Mackenzie River Break-up, Hydrologic Aspects of Northern Pipeline Development. Glaciology Division, Water Resources Branch. Department of the Environment, Ottawa, Canada, pp. 71–107.

McNamara, J.P., 2000. Bankfull Flow, Hydraulic Geometry, and River Ice in a Northern River. AWRA Water Resources in Extreme Environments, Anchorage, AK, pp. 191–196.

Payette, S., 1980. Les grandes crues glacielles de la rivière aux feuilles (Nouveau-Québec): une analyse dendrochronologique. Le Naturaliste Canadien 107, 215–225 (in French).

Prowse, T.D., 2001. River-ice ecology: hydrologic, geomorphic and water-quality aspects. Journal of Cold Regions Engineering 15, 1–16.

Prowse, T.D., Gridley, N.C., 1993. Environmental Aspects of River Ice. NHRI Science Report no. 5. National Hydrology Research Institute, Saskatoon, SK, Canada. 155 pp.

Rosgen, D., 1996. Applied River Morphology. Wildlife Hydrology, Pagosa Springs, CO. 390 pp.

Royer, A., Rousseau, A.N., Savary, S., Fortin, J.-P., Turcotte, R., 2006. PHYSITEL, un SIG pour la mise en place de modèles hydrologiques, 2e Symposium Scientifique Ouranos, Montréal.

Singh, V.P., 2003. On the theories of hydraulic geometry. International Journal of Sediment Research 18, 196–218.

Smith, D.G., 1979. Effects of channel enlargement by river ice processes on bankfull discharge in Alberta, Canada. Water Resources Research 15, 469–475.

Smith, D.G., 1980. River ice processes: thresholds and geomorphologic effects in northern and mountain rivers. In: Coates, D.R., Vitek, J.D. (Eds.), Thresholds in Geomorphology. Allen & Unwin, London, pp. 323–343.

Smith, D.G., Reynolds, D.M., 1983. Tree scars to determine the frequency and stage of high magnitude river ice drives and jams. Canadian Water Resources Journal 8, 77–94.

Smith, D.G., Pearce, C.M., 2002. Ice jams-caused fluvial gullies and scour holes on northern river flood plains. Geomorphology 42, 85–95.

Société d'Énergie de la Baie James, 1984, Aerial photographs, 1:10 000, Sept. 17, Lines 152 to 266.

Thorne, C.R., 1998. Stream Reconnaissance Handbook: Geomorphological Investigation and Analysis of River Channels. John Wiley, New York. 133 pp.

Thorne, C.R., Allen, R.G., Simon, A., 1996. Geomorphological river channel reconnaissance for river analysis, engineering and management. Transactions of the Institute of British Geographers 21, 469–483.

Vincent, J.S., 1985. Surficial Geology, Radisson, Québec, 1:100 000. Geological Survey of Canada. “A” Series Map 1591A.

Walker, J.H., Hudson, P.F., 2003. Hydrologic and geomorphic processes in the Colville River delta, Alaska. Geomorphology 56, 291–303.

Williams, G.P., 1978. Bank-full discharge of rivers. Water Resources Research 14, 1141–1154.

Wolman, M.G., Leopold, L.B., 1957. River floodplains: some observations on their formation. U.S. Geological Survey Professional Paper no. 282-C, Washington, DC, pp. 87–109.

Wolman, M.G., Miller, J.P., 1960. Magnitude and frequency of forces in geomorphic processes. Journal of Geology 68, 54–74.