



Ice-Related Floods and Flood Delineation along Streams and Small Rivers

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ABSTRACT: Streams and small rivers drain relatively small watersheds and are dominant in the headwater region of large drainage systems. Globally, their cumulative length is an order of magnitude superior to the length of larger rivers and, despite their limited width and depth, they can generate significant flooding and severe damage. Streams that have shaped, but also adapted to valleys of hilly topography present highly variable morphologies and hydraulic conditions over relatively short distances. A source and a consequence of this diversity is the great variety and relative intensity of ice processes taking place in their channel during the cold season. Most of these processes, passive or dynamic, can generate flooding under specific, but common hydrometeorological conditions. The rise in water levels along streams in hilly and mountainous regions (gradient above 0.1%) is not only associated with an increased channel roughness, but also with partial to complete channel blockage. Their unique winter hydrological regime deserves a special consideration in flood analyses and mitigation. This paper presents an overview of distinct ice processes taking place in small and steep streams and it illustrates, using quantitative examples, the preponderance of high water levels and flooding events during the cold season. Approaches to determine the frequency of ice-related floods are compared to those applied to larger rivers. This paper is relevant to river engineers and scientists working above 40 degrees of latitude. While it emphasizes flooding issues, it is also of interest to fluvial geomorphologists as well as aquatic and riparian biologists.

1. Introduction

A combination of the natural variability of hydrological processes, the impact of climate change, and human development in flood-prone areas has resulted in greater flood damages around the globe in recent years. In Canada, this has led to an accelerated governmental effort to delineate flood hazard areas along watercourses. Research and technical groups have been formed and funded to define state-of-the-art techniques to delineate flood hazard, to quantify flood risk, and to lessen flood damages.

Prior publications of this research group have emphasized that ice-jam floods should be considered in the delineation of the flood zone in cold regions (Burrell et al., 1995; Kovachis et al., 2017). This effort was inspired by the known fact that the frequency of ice jam-induced high water levels is often considerable and, at many locations, greater than the frequency of open water high water levels (Beltaos, 1995, Humes and Dublin, 1988, Tuthill et al., 1996). In addition to the flood elevation, ice jams are also associated with the presence of drifting ice floes on the flood plain, which increases the damage potential to infrastructure (Kovachis et al., 2017) and therefore affects what is defined as the global flood risk, i.e., the frequency of high water levels multiplied by the exposed vulnerability.

Small watercourses of highly variable sizes and morphologies dominate headwater regions of large drainage systems and small coastal watersheds. Despite their relatively limited width, depth and carrying discharge, they can generate significant flooding and severe damage. In cold countries, a broad diversity of intense ice processes often result in the blockage of stream and small river channels, especially if steep, causing water levels to rise above the bank elevation, despite the low discharge. These processes cause water and ice floes to spread onto the flood plain, and can lead to formation of ice (a few centimetres to many metres thick) outside the channel margin.

The present paper focuses on the winter dynamics of small watercourses, which are commonly referred to as streams and small rivers. The characteristics of small watercourses and ice processes are described in the next sections. This is followed by a discussion of the hydraulic effects of ice processes and of flood hazard and risk delineation. A positive collateral effect of this research is the potential development of improved methods to forecast and mitigate ice-induced floods. This paper provides useful information to engineers designing infrastructure along streams in areas above 40° latitude, as well as to public security services and other public agencies in charge of organizing concerted efforts to improve the societal resilience to floods.

2. Characteristics of streams

This paper focuses on ice-induced floods taking place along small rivers and streams, which will simply be referred to as “streams”. Table 1 presents a list of quantitative (hydrological and geometric) parameters that can be used to differentiate streams from smaller, ephemeral headwater channels and from larger rivers. These ranges largely correspond to what Meybeck et al. (1996) considers a small river, a stream, or a small stream, and it could include some brooks.

Table 1. Lower and upper limits of parameters that discriminate Streams from smaller or larger channels

Parameter	Lower limit	Upper limit
Drainage area	1 km ²	2000 km ²
Bankful width	1 m	100 m
Bankful depth	0.5 m	5 m
Bankful discharge	1 m ³ /s	1000 m ³ /s
Channel gradient	0.1%	NA
Stream order	2	6

Beyond those parameters, a stream can present different patterns and profiles that can vary over relatively short distances. Possible areal patterns include linear channels, wandering channels, braided channels, and occasional anastomosed reaches. In turn, meandering channels are normally associated with gradients below 0.1%. Based on the channel classification proposed by Montgomery and Buffington (1997), typical stream profiles include cascades, step-pools, rapids, and riffle-pools whereas watercourses with channel beds featuring dunes or ripples are normally excluded from what is considered a stream.

A publication by Turcotte and Morse (2013) presented a conceptual model, in the form of a graph or a table, that makes a link between the morphology of a reach, or its gradient, and typical ice processes that can occur along that reach, considering its size and the prevailing climate. This model, along with other morphology-based models (e.g., Bergeron et al., 2011), can be used as a first step to identify potential ice processes that can take place along stream reaches of specific physical characteristics. Morphological studies of various watercourses (Cao Shu-you, 2002; Montgomery and Buffington, 1997) indicate that there is a broad tendency of channel slope to increase as channel size decreases. This however, is not always the case, especially along heterogeneous geologies or in post-glacial regions where suspended valleys are common.

3. Qualitative characteristics of stream ice processes

In large channels, most ice-induced flooding episodes are caused by hydro-mechanical events associated with the rupture of an ice cover as the discharge increases (i.e., breakup ice jams; e.g., Beltaos, 2008), a process mostly associated with melting meteorological conditions, but also flow regulation, or northward watershed orientation. To a lesser extent, large channels can also be flooded by dynamic or passive hydro-thermal freeze-up processes taking place under cold conditions (Beltaos, 2013). These large-channel ice processes are different types of an ice jam, defined as “a stationary accumulation of fragmented ice or frazil that restricts flow”, by IAHR (1986).

Ice jams of different types also occur in streams, including very small channels (e.g., Turcotte et al., 2012a). However, in streams, thermally driven ice processes, such as ice dam formation, may represent a dominant source of ice-induced floods, which can take place under very cold conditions. This section introduces and qualitatively describes these stream ice processes.

3.1 Anchor ice

Under super-cooled conditions in turbulent streams, anchor ice initially forms on the channel bed by entrainment of frazil particles or from direct nucleation of ice crystals on substrate (e.g.,

Malenchak and Clark, 2013), and it further thickens by in situ growth or by interception of drifting frazil particles (Figure 1A). Anchor ice is generally seen on top of the substrate, but it can also be (hydraulically) compacted within the substrate in shallow and/or high turbulent areas (e.g., Stickler and Alfredsen, 2009). A flume experiment by Kerr et al. (2002) shows how the roughness of the anchor ice-covered bed may initially increase and then decrease as the surface of the anchor ice becomes smoothed. The presence of anchor ice also raises the effective bed elevation and this condition, together with roughness alteration, reduces the flow conveyance capacity of the stream, generating a rise in water levels, with potential local flooding, especially if the anchor ice layer develops over many days (e.g., Stickler et al., 2010).

Asvall (1997) described a combined anchor ice- and frazil ice- flooding event that affected multiple basements and led to a shutdown of the sewage system in the city of Otta (Norway). Strømslid et al. (2012) reported that anchor ice formation and frazil accumulation in the Bardu River (Norway) can generate significant local flooding, affecting roads and farmlands. They further mentioned that the morning release of anchor ice (which is part of the most common anchor ice diurnal cycle reported in the literature) contributed in downstream ice jamming that generated further flooding (a process that is often reported in the literature and described further on). In this particular case, hydropower regulation exacerbated the impact. In turn, Girling and Groenvelde (1999) reported an anchor ice-induced water level rise that reduced the head of the Limestone generation station (Canada), leading to an annual production loss of over one million dollars. On a more positive note, anchor ice-induced floods along streams may be beneficial to the riparian vegetation (e.g., Lind and Nilsson, 2015).

3.2 Ice dams

Anchor ice development is not always spatially homogeneous and surface smoothing may never occur. At steps in the channel profile or where large rocks are aligned across more uniform channel profiles, anchor ice development may be locally promoted to the extent where it becomes emergent (Figure 1B). Ice dams form where anchor ice creates a channel blockage and this initiates the process of icing, which is the progressive flooding and freezing of water on cold surfaces (see also section 3.4). As the water flows on the dam surface, more ice is produced and the upstream water level further rises, which can eventually produce overbank flooding. This freeze-up process is described in more detail by Turcotte et al. (2011, 2013).

Perforated or breached ice dams can remain in place throughout the winter, letting low flows freely transit underneath. Though breached ice dams represent relatively fragile structures, they can resist increasing runoff long enough to generate overbank flooding, especially in narrow channels where high dams are frozen against (or supported by) the bank or in the presence of a thick, stabilizing, in-channel snow layer. The structural failure of a single ice dam under an increasing discharge can generate a cascade effect leading to a dynamic breakup over a significant downstream distance with a potentially damaging ice run and ice jam. Observations from Scandinavia and field data from the Montmorency River (Canada, Figure 2), suggest that the sudden failure of series of ice dams can even take place during their formation, under cold conditions and a stable, low runoff. The wave that results from the failure of ice dams, either under cold conditions or an increasing discharge, can be termed “dave” for “ice dam released wave”. Only time will tell if this “cold breakup” process is common enough for this terminology to be adopted by the river ice scientific community.

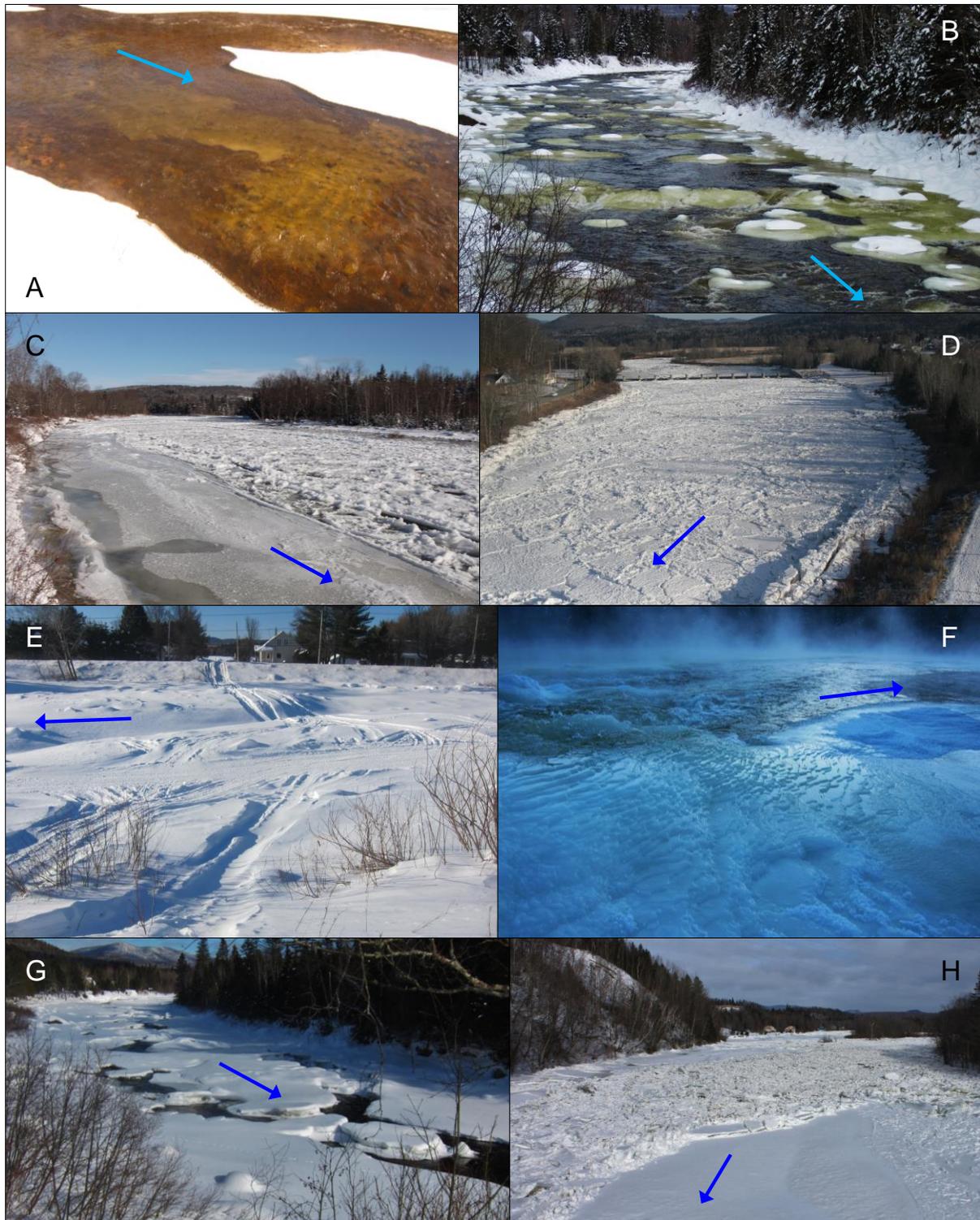


Figure 1. (A) Anchor ice patches in riffle-pool reach of the Montmorency River, QC, Canada; (B) Ice dam across the rapids reach of the Montmorency River; (C) Frazil jam in a riffle-pool reach of the Ste. Anne River, QC, Canada; (D) Grounded hanging dam in the Ste. Anne River; (E) Undulating ice surface looking across the Ste. Anne hanging dam during winter; (F) Icing layers in the rapids of the Montmorency River; (G) Partial ice cover at the end of winter in the Montmorency River (Same site as B); (H) Ice jam against an intact ice cover in the anastomosed reach of the Montmorency River.

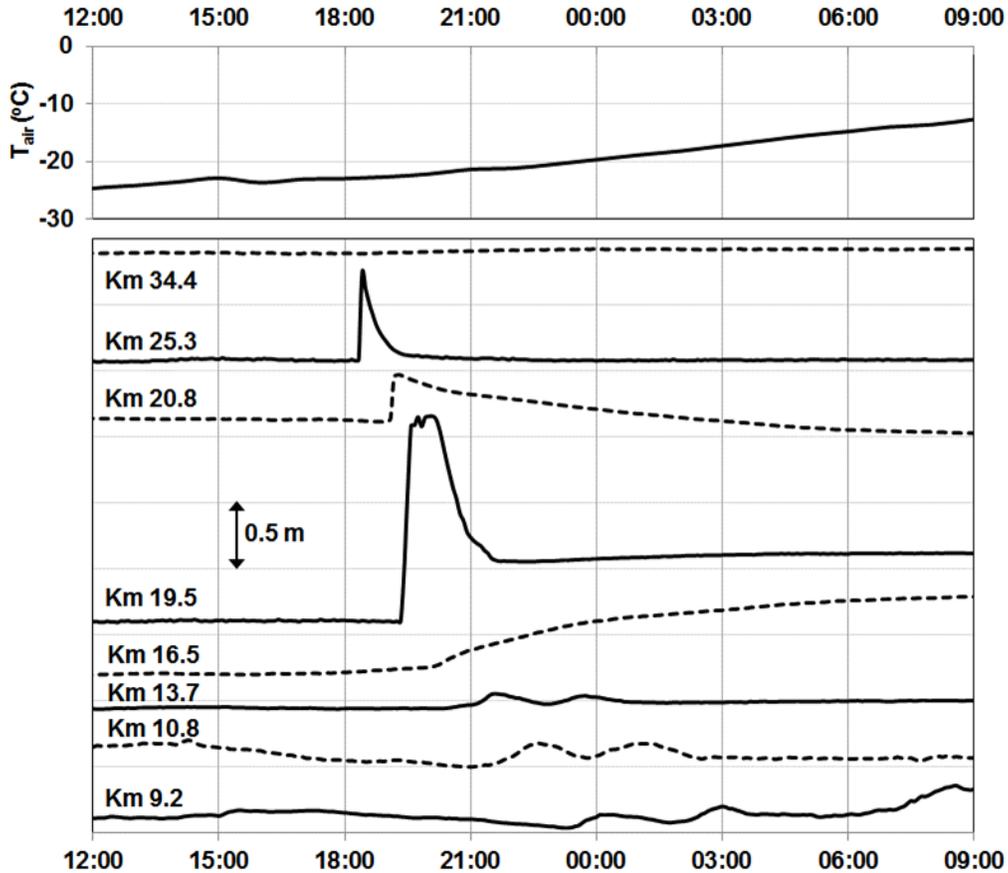


Figure 2. Cold breakup (between km 25.3 and 34.4), ice jam (downstream of km 19.5) and residual wave (downstream of km 16.5) signals along the Montmorency River, QC, Canada, from Jan 8th to Jan. 9th, 2015.

3.3 Freeze-up jams

3.1.1 Surface juxtaposition

A simple surface juxtaposition of ice pans and floes within, or downstream of, steep stream reaches is improbable because (1) turbulent conditions contribute to maintaining drifting ice in the form of unconsolidated slush, even under cold conditions; (2) because waves and rocks readily break thin drifting ice floes in small pieces; and (3) high surface velocities would lead to the formation of thickened jams (next subsection). In turn, in flatter reaches, surface juxtaposition of ice floes may occur, but it is unlikely to produce any overbank flooding.

3.3.2 Thickened jam (dynamic consolidation)

In streams, thickened freeze-up jams are often the result of frazil interception and packing under cold conditions. For example, along the Ste. Anne and Montmorency Rivers (Canada) over several winters, riffle-pool reaches with a gradient below 0.3% can be dynamically covered by a frazil jam (Figure 1C) that migrates in the upstream direction whereas rapids with a gradient above 0.5% can hardly support such a process, apparently independently of the air temperature and discharge.

More commonly along larger rivers, but also in streams, freeze-up jams are associated with ice runs intercepted by a stable surface ice cover section, which would be the case downstream of steeper reaches. These ice runs can either originate from an unstable ice cover development (or hydrological) condition or from an increasing runoff that lifts, breaks, and carries the recently formed ice cover. Since the ice cover is often thin in early winter, the probability of a “wide channel jam” made by shoving and compaction (e.g., Beltaos, 1995) is higher for a given stream during the first winter cold spells. In turn, a “narrow channel jam” is less likely to produce flooding because of its relatively limited thickness. Depending on the prevailing hydrological and hydraulic conditions, thickened freeze-up jams made of frazil or ice floes can generate overbank flooding in a similar way as described in section 3.5 for breakup ice jams.

3.3.3. Hanging dams (passive accumulation)

A surface ice cover usually forms first in low gradient segments such as pools and lakes or reservoir heads while frazil production (and snow slush entrainment) in steeper upstream reaches can last several additional days or weeks. In these circumstances, if the drifting frazil slush or ice floes do not accumulate in the form of a thickened ice jam, they will be entrained under the ice cover. The river-ice literature suggests that a surface velocity above 0.7 m/s will lead to ice floes entrainment whereas unconsolidated frazil slush may be entrained under a floating obstacle at much lower surface velocities (Beltaos, 2013). Such ice front dynamics also depend on prevailing air temperature, supercooling conditions, and channel depth.

In streams, entrained ice floes may become grounded, and in this case, a thickened jam would result. Similarly, large portions of a hanging dam made of frazil and snow slush may become grounded (e.g., Vergeynst et al., 2017) and its surface appearance may resemble that of a thickened ice jam (Figure 1D). As the ice accumulates under the ice cover, upstream velocities may be reduced enough to promote the frontal (upstream) progression of a surface or a thickened jam. Therefore, the ice cover formation may alternate between dynamic frontal progression and passive downstream deposition and thickening. Once most of the river is covered, the combination of a receding discharge and channel insulation contributes to thermally erode a tunnel into the hanging dam core. This causes the hanging dam to become increasingly grounded, as can be presumed by observing its undulated surface (Figure 1E).

In unregulated streams, the reported presence of a hanging dam may indicate a recurrent ice accumulation location whereas in regulated streams, the presence, location and thickness of a hanging dam may vary significantly from one winter to another. Under specific hydrometeorological conditions affecting ice deposition patterns and thicknesses, overbank flooding and icing may occur. Floods caused by hanging dams are relatively rare, but have been reported on small rivers (e.g. Beltaos et al., 2007a; 2007b; Vergeynst et al., 2017). A large hanging dam in a channel represents an important hydraulic and drifting ice obstacle, and it takes a significant amount of heat or rise in discharge to respectively melt it or mobilize it.

3.4 Aufeis and icing

Aufeis formation is mostly known as a post-freeze-up process occurring in braided channels of arctic and subarctic regions (Grayson, 2010, Daly, 2013). Aufeis represents an ice accumulation formed by progressive freezing of water layers that emerge from under an existing ice cover through cracks because of pressurized flow conditions. This may originate from a complex,

spatiotemporal combination of (1) ice-induced flow fluctuations and (2) a progressive reduction in the water carrying capacity of the channel. Aufeis is known to generate extensive overbank ice production under cold conditions, and it can also generate additional flooding as the discharge increases in the spring while the channel is still blocked by ice (e.g., Daly et al., 2011; Toniolo et al., 2017).

More generally, and including braided channel aufeis, stream “icing” is the general process of ice layers forming on cold surfaces including banks, in-channel emerging rocks, and existing ice surfaces. This process follows anchor ice development, it represents an important part of ice dams’ formation (e.g., Turcotte et al., 2011) and it is therefore often associated with ice development in steep streams (Figure 1F). Culverts, especially those made of steel, are prone to generate icing. This process can also be observed where groundwater seepage or tributary flows cannot find their way under an existing ice cover. Carey (1973) contains additional information on stream icing.

3.5 Breakup ice jams

Breakup ice jams in sizeable rivers are almost always of the wide-channel kind. As the channel size decreases, narrow-channel jams may be more common. Criteria for distinguishing between the two types of jams are discussed in Beltaos (1995). The most common hydrological condition that leads to breakup ice jams is when the runoff increases, either in the spring or during winter in temperate regions, but breakup ice jams can also be caused by reservoir management. Hydropower operation can be restricted in winter to avoid river ice breakup (e.g., Gebre et al. 2013).

In low gradient streams, where a floating surface ice cover is present, the increasing flow and stage lift the cover and separate it from bank and bed locations of attachment. Eventually, the cover is mobilized and soon afterwards breaks down into ice blocks. In steeper reaches, the ice cover is more fragile because it is often initially fragmented (e.g., Turcotte et al., 2012b; Figure 1G). In this case, the increasing discharge can mobilize small ice blocks that may become trapped against rocks and thicker ice slabs. This blockage generates a local stage rise and an increased stress that can result in a reach-scale ice cover mobilisation. In all cases, the resulting ice run may be arrested by intact ice farther downstream and form a jam.

Different hydraulic, morphologic and cryologic conditions promote ice jamming and several authors have been able to specify channel characteristics where ice jams are common. Typical ice jam locations that are specifically attributed to streams, based on the writers’ experience, include:

- At slope reductions, including large pools, reservoir or lake entrances,
- Entering multichannel reaches (secondary or anastomosed channels that evacuate water),
- At locations of low and wide flood plains (becoming a floodway; e.g., Calkins, 1983),
- At sharp bends and constrictions,
- Downstream of confluences (especially when tributaries breakup first),
- At channel enlargements (where the ice run energy is dissipated over a large surface), and
- Against large obstacles such as boulders, culverts or in-channel structures.

In most cases, the presence of an intact ice cover is required to initiate ice jamming (Figure 1H). Therefore, apart from reaches located downstream of lakes (e.g., Lind et al., 2016) or large reservoirs, ice jams can take place along any stream sizes and morphologies (Turcotte and Morse, 2013).

Breakup jams sometimes contain not only ice blocks but also frazil slush. The latter originates in relatively thick freeze-up jams that never froze solid during the winter. This is especially likely to occur in streams (e.g., Pigeon et al., 2015) or even in larger rivers that contain rapids (a good example is the relatively large Athabasca River above Fort McMurray along which the average slope is nearly 0.1%). The presence of frazil and snow slush can reduce seepage through the voids inside the jam (Beltaos, 1999) and generate higher upstream water levels.

4. Hydraulic impact of stream ice processes

This section presents water depth and discharge data associated with various stream ice processes and illustrates their importance to delineate the flood zone.

4.1 Anchor ice and ice dams

In cascades and step-pool streams, often of the lowest channel order, the very initial phase of anchor ice development seems heterogeneous in space and the location where ice dams will develop becomes immediately obvious. Ice dams as high as 4 m have been reported in the literature (Strømshlid et al., 2012; Tesaker, 1996). Figure 3 presents a discharge and water depth data set of Lépine Creek (step-pool channel, watershed area of 7 km², average maximum discharge of 8 m³/s), from the Montmorency River watershed (Canada) for winter 2011-2012 (considered here as a short winter with very few mid-winter runoff events). Clearly, ice dams can generate overbank flooding at very low discharges in that channel (as synthesized by Figure 3A). During a midwinter or an early spring runoff event, ice dams can support relatively high pressures and can generate further flooding (Figure 3B). Based on multiple winters of survey along that stream, the right graph of Figure 3 also suggests that ice dams are generally mobilized and evacuated by a discharge well-below the equivalent open water bankfull discharge (Q₂). However, this may not be the case along all steep streams and significant overbank flooding could occur if ice dams are more structurally resistant.

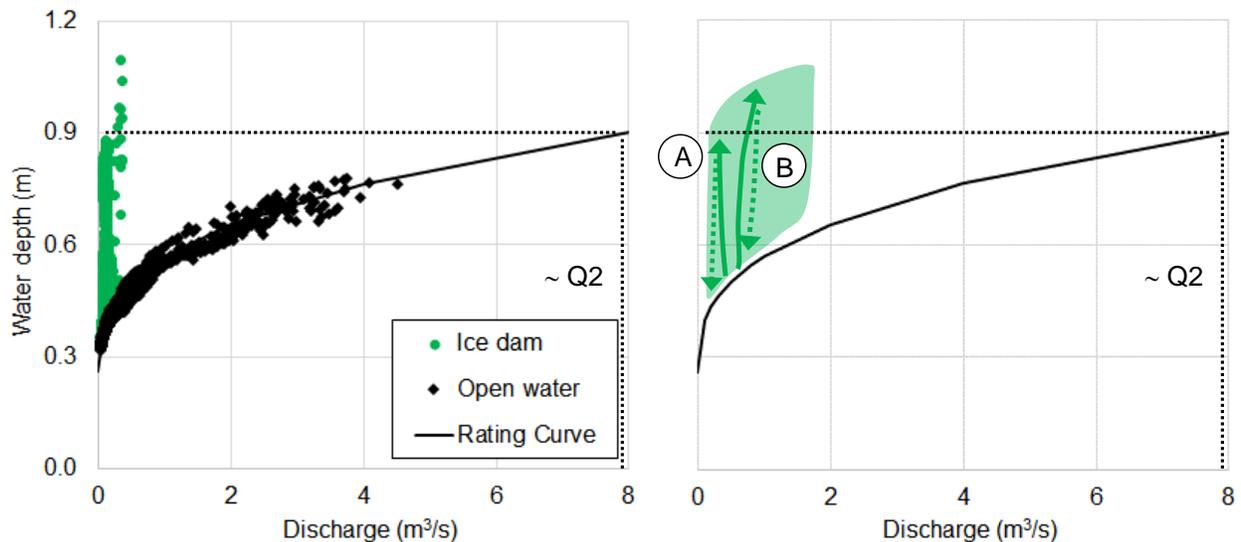


Figure 3. Hydraulic impact of ice dams as illustrated by an hourly data set from Lépine Creek (step-pool channel draining about 7 km²), Canada, from November 2011 to June 2012 and by a conceptual graph showing the path (1) of the initial ice dam formation and (2) of a runoff event in the presence of breached ice dams and snow.

In turn, along rapids or riffles, and more generally in wider and deeper streams, homogeneous anchor ice blankets form first and can eventually evolve into well-defined ice dams, a transition that can take several days to weeks, depending on multiple morphologic, groundwater and hydrometeorological factors. Figure 4 presents hourly discharge and water depth data of the Montmorency River (rapids, watershed area of 1100 km², average maximum discharge of 440 m³/s, Figure 1B), Canada, from November 2011 to June 2012, and distinguishes three freeze-up and mid-winter ice processes. The right graph in Figure 4 presents the potential hydraulic impact of (A) a developing anchor ice blanket, (B) the formation and breaching of ice dams, and (C) breached ice dams and suspended ice cover as the discharge rises. In wide streams, ice dams are more likely to breach for small increments in discharge, but they can temporarily store a significant quantity of water upstream, which exacerbates an eventual dynamic breakup event. Overall the ice cover in rapids is fragile (Figure 1G) and should be flushed at a discharge well-below the equivalent bankfull discharge. Along riffle pool channels, the evacuation of the fragmented ice cover can be impeded by the presence of a thick surface ice cover with possible hanging dams (Bergeron et al., 2011).

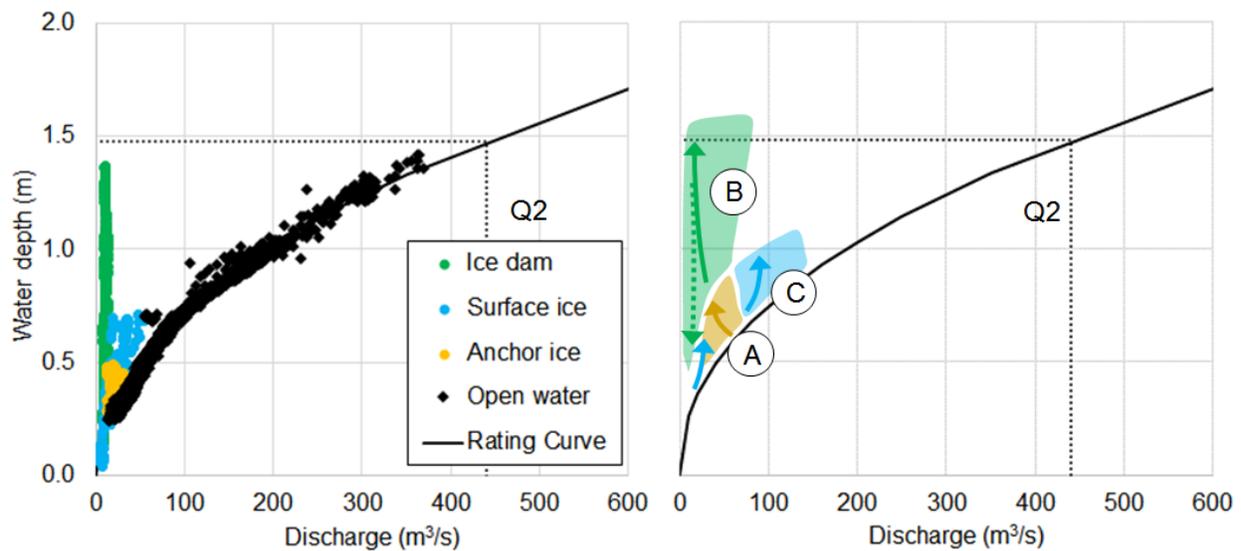


Figure 4. Hydraulic impact of anchor ice, ice dams and surface ice as illustrated by an hourly data set from the Montmorency River (rapids draining about 1100 km²), Canada, from November 2011 to June 2012 and by a conceptual graph showing the potential discharge – depth relationship for (A) anchor ice, (B) ice dam development and (C) a partially suspended surface ice cover.

Finally, Figure 5 presents the hydraulic impact of the sporadic and temporary formation of anchor ice blankets at the Syrstad gauge in the regulated Orkla River in Norway (considered at the upper limit of a stream with a watershed area of 2280 km²) from October 2010 to April 2011. This phenomenon does not occur at the same intensity every winter, as opposed to what is presented in Figures 3 and 4. The development of anchor ice blankets is not often associated with overbank flooding and under an increasing runoff and warmer air temperature, the blanket should be released from the bed before a significant stage rise occurs. It has been recently found that the release of anchor ice is, at least occasionally, hydraulically driven (Jasek et al., 2015).

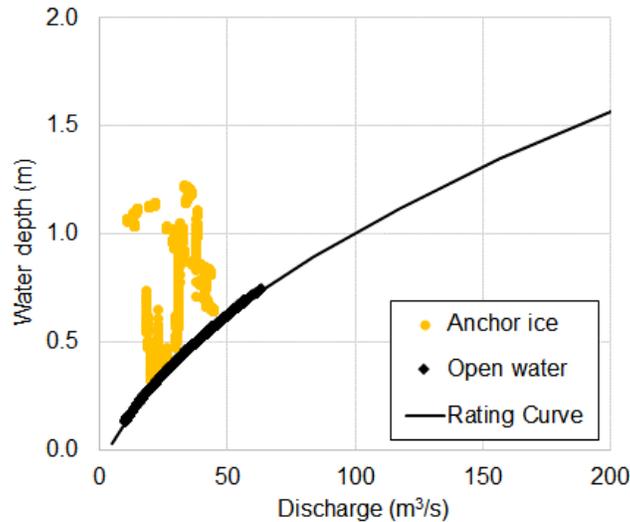


Figure 5. Hydraulic impact of anchor ice blankets as illustrated by an hourly data set from the regulated Orkla River (riffle-pool channel draining about 2280 km²), Norway, from October 2010 to April 2011. The data is corrected for ice presence on a 24 hour basis and should be considered as an approximation.

There is a need for additional research that could reveal how a stream cross section geometry and channel profile may affect the resistance and buildup rates of ice dams. While buildup rates as high as 10 cm/h have been reported, the most common buildup rates for streams range from 5 mm/h to 20 mm/h (e.g., Dubé et al., 2015). This rate normally endures as long as there is heat loss from the channel to the atmosphere. Therefore, for a cold spell of given duration and intensity, overbank flooding caused by ice dams is more likely to occur along shallow channels. This means that for a specific region or watershed, ice-dam-induced floods are more likely to occur first along low order channels (2 to 4). This however may not apply to wide braided channels where the freeze-up discharge is small compared with the overall capacity of the channel.

4.2 Thickened freeze-up jam (dynamic consolidation)

Andres et al. (2005) presented an impressive freeze-up jam data set for the Peace River (a large channel) that illustrates the potential impact of early winter consolidation events. These jams are of the wide-channel kind and their formation and flooding potential are often linked to large fall and winter flows generated by regulation (Neill and Andres, 1984).

Freeze-up jams along streams have not been extensively documented. Figure 6 presents a data set from a riffle-pool reach of the Ste. Anne River (watershed area of 700 km², average maximum discharge of 280 m³/s) for winter 2015-2016 at a location (Figure 1C) where a freeze-up jam largely made of frazil was observed during three consecutive winters. A first frazil packing event led to slight overbank flooding (Figure 6A). Prior to breakup, the frazil and ice jam resisted a significant discharge increase and a portion of the water was evacuated in the adjacent forest and fields (Figure 6C). Another potential threatening situation is when a runoff event occurs just after the formation of a freeze-up jam, when the ice cover has not been thermally eroded and when its structural resistance is already significant (Figure 6B). The intensity of ice processes in 2015-2016 at that site, as presented on the left side of Figure 6, were probably above average, but such processes seem to generate local overbank flooding more frequently than open water runoff events.

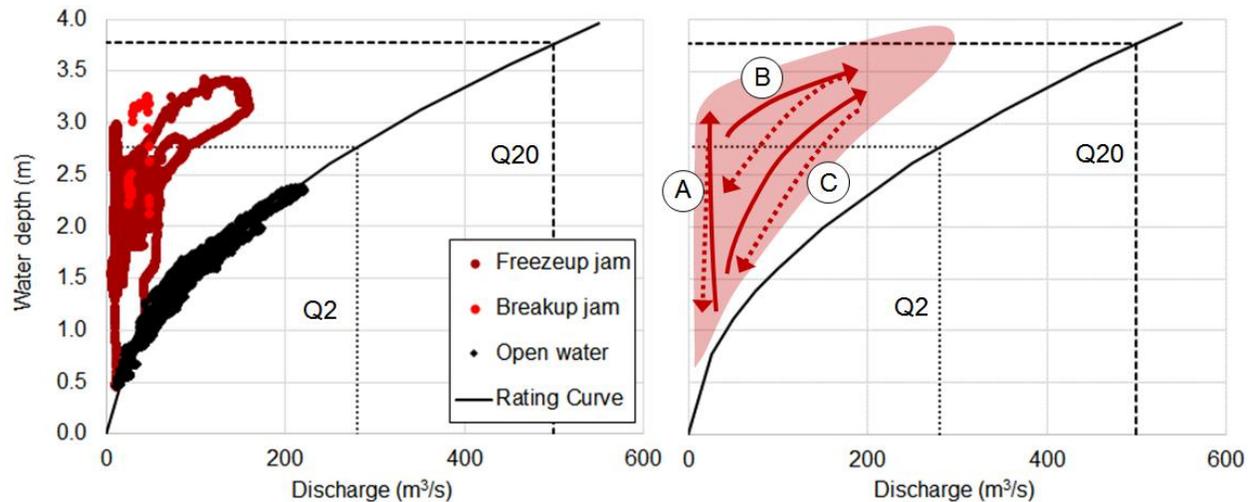


Figure 6. Hydraulic impact of a freeze-up jam and its remnants as illustrated by a data set from the Ste. Anne River (riffle-pool channel draining about 700 km²) at a rate of one measurement every 7.5 minutes from November 2015 to April 2016 and by a conceptual graph showing the potential discharge – depth relationship for (A) the initial consolidation event, (B) an early winter runoff event and (C) a late winter runoff event.

The right hand side of Figure 6 also suggests that a freeze-up jam may remain in place at very high discharges. In that case, most of the water would be evacuated through the flood plain, an eventuality that is more likely to occur if the freeze-up jam has formed where a low flood plain or secondary channels exist. Knowing the maximum discharge that can be supported by the toe of an ice jam (ice clearing discharge; Beltaos, 2008) is important to understand its flooding potential and frequency; unfortunately, this information is particularly difficult to evaluate or to simulate.

4.3 Hanging dams (passive accumulation)

Quantitative data sets of hanging dams in streams have been presented by Allard (2011), Beltaos (2007a,b), and Vergeynst et al. (2017). The authors have commented on the potentially important frazil accumulation thickness (4 m in Vergeynst et al., 2017) and on its spatial heterogeneity. As opposed to consolidated ice jams, and because of the physics of passive frazil deposition (Beltaos, 2013), their initial roughness is normally relatively low and comparable to that of the bed. However, their gradual development over several days or weeks can generate a significant backwater impact because of channel blockage, especially in shallow stream segments located downstream of steep reaches or of a flow-regulating dam. Overbank flooding can occur gradually during cold spells (frazil production) or more suddenly if the discharge increases (runoff or regulation).

The left side of Figure 7 presents a data set obtained from the Ste. Anne River at a site (Figure 1D) that has been surveyed by Vergeynst et al., (2017). Water levels were measured 8 times per hour from November 2015 to April 2016. The water level achieved in 2015-2016 was higher than during the previous and following winters, but was significantly below the hanging dam induced flood of winter 2003-2004.

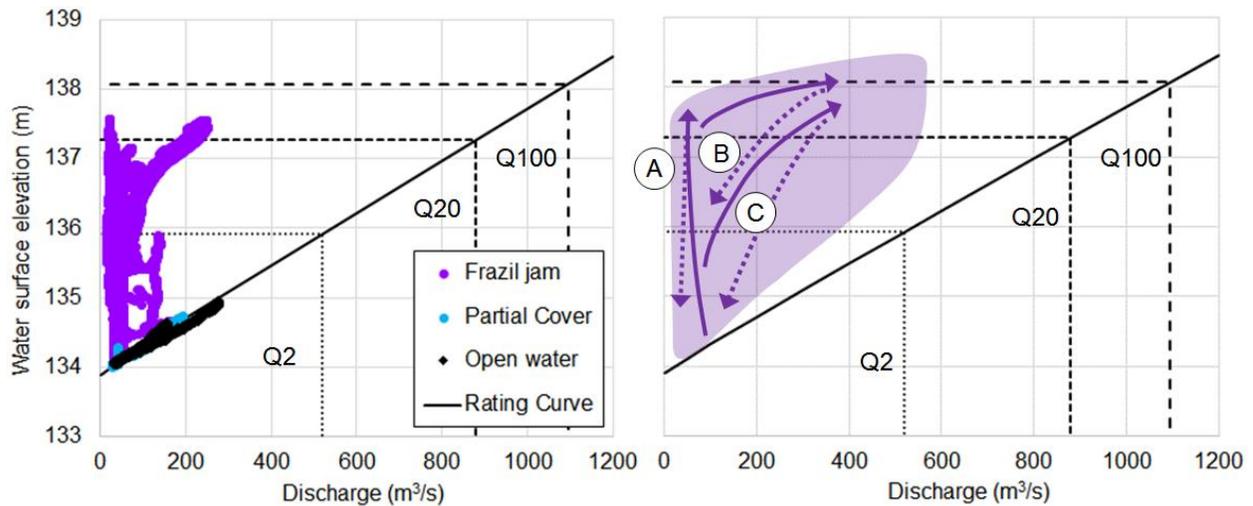


Figure 7. Hydraulic impact of a hanging dam as illustrated by a data set from the Ste. Anne River (riffle-pool channel draining about 700 km²) at a rate of one measurement every 7.5 minutes from November 2015 to April 2016 and by a conceptual graph showing the potential discharge – depth relationship for (A) the initial development event, (B) an early winter runoff event and (C) a late winter runoff event.

The initial frazil accumulation process in the Ste. Anne River usually begins as the early winter discharge drops (Figure 7A) and this process continues as long as upstream frazil production reaches are not covered more than 50% (Turcotte and Morse, 2016), which can take more than one month. The most threatening timing for a winter runoff event, from a flooding point of view, occurs immediately after the hanging dam has reached its maximum thickness, because this brings more ice from upstream reaches and generates further blockage of the channel (Figure 7B). A mid-winter or early spring runoff event can also cause overbank flooding (Figure 7C). The combined ice-hydraulic behaviour presented on the right side of Figure 7 could be expected along most deep stream sections located downstream of long stretches of rapids or waterfalls as well as in regulated systems. Of course, the values presented on the axes (including Q2, Q20 and Q100) are site specific.

Two threatening aspects of hanging dams, from a flooding perspective, are (1) that they contain a significant amount of heat deficit (in the order of 50 kWh/m³), which can prolong their presence in the spring and (2) that they can intercept any incoming drifting ice floes or major ice runs and initiate a breakup jam. On a more positive note, a possible open-water reach upstream can carry an increasing amount of heat during the second half of winter, which can initiate melting well before the snowmelt runoff period.

4.4 Aufeis

Development processes of aufeis in arctic and sub-arctic regions are currently not well-understood and very few quantitative data related to aufeis have been collected (with no known continuous, winter-long, ice thickness or water level measurements). The global understanding is that their formation can begin as soon as the entire channel is ice-covered and that they can thicken until late winter, which in the arctic corresponds to the month of May. Aufeis many metres in thickness have been reported in northern Asia and North America (e.g., Sloan, 1976; Toniolo et al., 2017). A case study of overbank flooding due to aufeis is described by Daly (2013).

Figure 8 proposes that aufeis forms when the discharge is relatively low (A) in streams where runoff may become depleted. It is uncertain if the water level (in terms of pressure) varies significantly between two development (ice-surface flooding) episodes. In turn, it is probable that their presence generates, and originates from, discharge variations. In the spring, the snowmelt runoff may dominantly flow on the ice surface (Figure 8B). Therefore, if the aufeis has already reached the flood plain level during the winter period, the spring freshet is likely to generate additional flooding (e.g., Toniolo et al., 2017). Eventually, an ice free channel may be melted or broken within the aufeis field (Figure 8C), but its flow carrying capacity may remain well below that of the entire braided channel for several additional weeks.

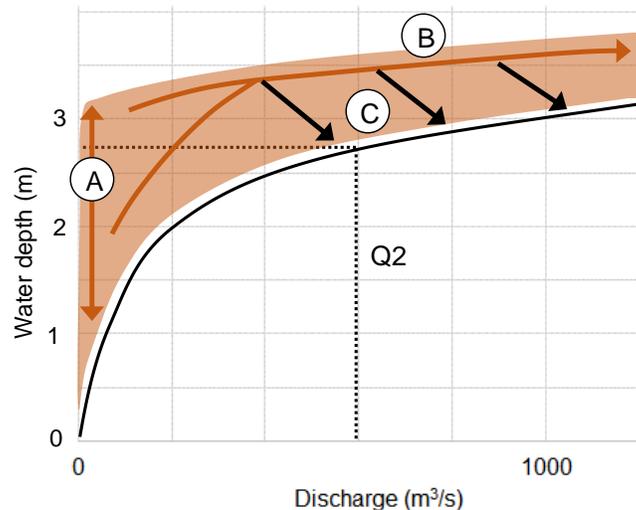


Figure 8. Conceptual graph showing the potential hydraulic impact of Aufeis in wide and shallow streams for (A) the initial (often winter-long) formation, (B) the initial spring runoff event that can lead to (C) the eventual breaching of the accumulation. This graph suggests that major aufeis accumulations can probably not be mobilized by high discharges if no significant melting occurs.

4.5 Breakup ice jams

Breakup ice jams represent the most common ice process associated with ice-induced floods. Water level data sets in the presence of an ice jam are common in large rivers and many events have also been monitored along streams. The left side of Figure 9 presents a data set that was obtained from the Ste. Anne River, at the foot of rapids and upstream of the site for which data is presented in Figure 6. The presence of a freeze-up jam a few kilometres downstream represented an ideal ice cover setting for the subsequent formation of a breakup jam. On April 1st, 2016, the fragile ice cover over 5 km of rapids was mobilized and eventually ran into the freeze-up jam, locally generating water levels in the order of the 100 year open water flood.

As the discharge rises, the ice cover over an increasingly large channel area will be mobilized. As a result, ice runs and ice jam formation and mobilisation events may result in the formation of increasingly large and long ice jams (as mentioned by She et al., 2009 for a large river system). This especially applies to streams presenting a variable morphology and gradient, and when the spring runoff event is divided into several peaks. The right hand side of Figure 9 indicates that ice-jam induced water levels generally increase suddenly (A) and may continue to rise gradually with the increasing discharge. In turn, water levels may drop gradually (B) if runoff decreases and if the jam slowly melts in place, or suddenly if the jam is mobilized and washed downstream (C).

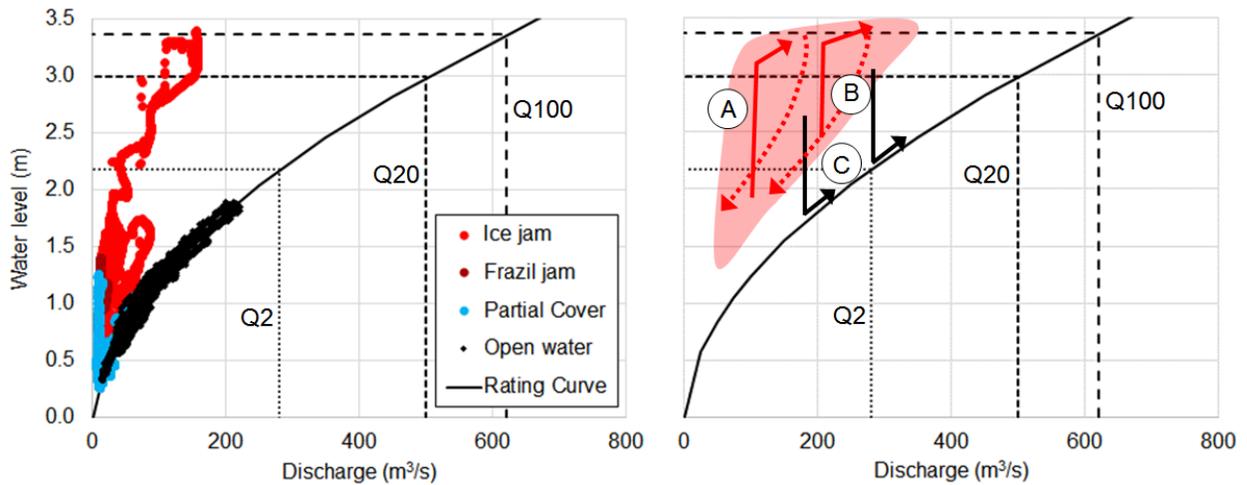


Figure 9. Hydraulic impact of a breakup ice jam as illustrated by a data set from the Ste. Anne River (rapids ending in a riffle-pool channel draining about 700 km²) at a rate of one measurement every 7.5 minutes from November 2015 to April 2016 and by a conceptual graph showing the potential discharge – depth relationship for an ice jam (A) forming and (B) remaining in place for the entire runoff event or (C) released and washed downstream of the hydraulic control that defines the open water rating curve.

Channel characteristics and hydraulic conditions in streams and small rivers are largely different from what is observed along larger rivers:

- The width-to-average-depth ratio is smaller and it cannot be assumed that the hydraulic radius is almost equal to the mean flow depth.
- The ratio of ice thickness (t) compared to channel width (W) or depth (H) is larger.
- The channel width (W), channel gradient (S_o) and channel depth (H) can vary significantly over short distances.
- The open water average velocity may well exceed 2 m/s (e.g., in rapids) and supercritical flow conditions are common, especially at high flows.

Figure 10 presents a dataset of breakup ice jams that have been documented along different streams expressed in the form of a ξ - η relationship, as introduced by Beltaos (1983). For all the documented jams, S_o is taken to be the average reach gradient (varying from 0.23 to 4%) where H was measured or estimated along the jam (from 1.4 to 4.0 m), W is considered to be the average width of the channel between shear walls remaining after the mobilisation or melting of the jam (varying from 5 to 60 m), the unit discharge q is equal to the total estimated water discharge Q (varying from 3 to 500 m³/s) divided by W , and g is the gravity constant.

The first data set (Figure 10, red) is based on the total discharge Q estimated upstream or downstream of the ice jam site whereas the second data set (Figure 10, blue) acknowledges that a certain portion of the discharge was flowing on the flood plain or in secondary channels (Q_f), as it was observed. This reduces the under-ice Q , which controls both components of H (submerged ice jam thickness plus under-ice flow depth).

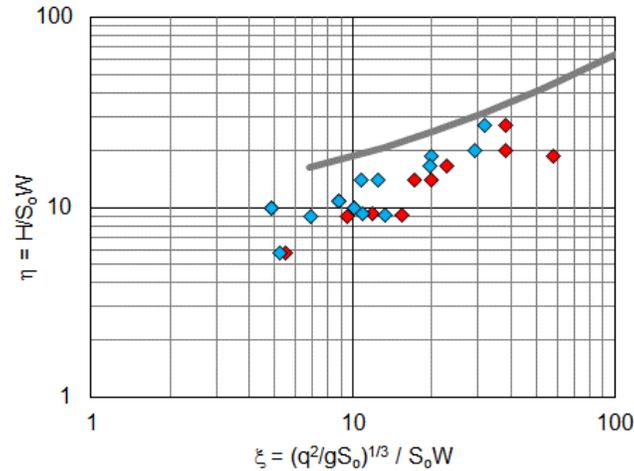


Figure 10. Data set of 14 distinct breakup ice jams documented along streams of various characteristics presented on a ξ - η graph. The empirical relationship inferred from the data presented by Beltaos (1983) for equilibrium wide-channel jams is shown as a thick grey line. Red diamonds are based on the total channel water discharge whereas blue diamonds are based on a reduced water discharge, considering overbank or secondary channel evacuation.

Results show that all data points plot below the empirical equilibrium jam relationship. Even when considering a reduced, more realistic main channel discharge ($Q-Q_f$), this relationship overestimates H by 1.2 to 2.4 times. Small deviations from the empirical line of Figure 10 could be explained by differences in internal friction and roughness characteristics of different wide-channel jams or by non-equilibrium conditions. This could apply to the six blue data points with $\eta \sim 15$ or more. Large deviations, however, point to jams of different type. Using Eq. 4.46 of Beltaos (1995), it can be shown that the thickness-related component of η must be at least equal to ~ 10 for equilibrium wide-channel jams. Consequently, all data points in Figure 10 for which η is close to, or under 10, describe jams of different type. These considerations yield the following hypotheses about breakup ice jams in streams:

- In small streams, ice jams may be of the narrow type, which causes lower water levels than the wide channel type. This is the case for reported ice jams observed in streams presenting a gradient superior to 1% or so.
- In shallow streams, ice jams may be largely grounded (the contact between large ice blocks and the channel bed represents an important, if not dominant, resisting force) and a proper equilibrium section may not be achieved, even if the channel is wide.
- In steep channels, Q_2 is evacuated at a high flow velocity and relatively low flow depth. Therefore, the presence of an ice jam (of potentially low porosity) immediately generates a significant flow velocity reduction and a proportional stage rise. This would explain why, for 11 of the 14 documented jams in Figure 10, the water level reached the flood plain.

It can be concluded that while jams in streams form at locations that compare to those of large rivers, they are more likely to remain in place where the flood plain is low or where secondary channels are present because this releases the pressure on the ice accumulation and promotes grounding. As a consequence, flood plains and secondary channels impose a limit to the jam thickness and maximum water level H , a finding also shared by Calkins (1983). Of course, the same applies to large rivers, but the top of the bank is not as readily attained as in streams and small rivers when ice jams are present.

5. Flood delineation along streams

Figures 3 to 9 explain some of the physics associated with specific ice processes, but they do not include any historical reference that enables the evaluation of the frequency of specific ice-induced floods and associated water levels. This section presents a number of approaches that can be applied to determine the frequency of ice-induced floods along streams and small rivers.

5.1 Direct stage-frequency method

The most precise and representative method to determine the frequency of high freeze-up, mid-winter and breakup water levels is to use a long record of water levels measured at the site of interest together with local ice observations. This type of data set is seldom available. For example, an aufeis-induced flood frequency analysis appears difficult to perform because aufeis are not well documented where there is no vulnerability whereas mitigation measures are often applied where they generate damage. Therefore, other approaches need to be considered.

Communities that are frequently flooded during the ice season should consider installing georeferenced water level loggers at different locations along their watercourses, an investment that represents a negligible fraction of the annual budget of any small town (and governmental public security services that deal with flood consequences). At least, communities should keep a record, either in terms of photographs or notes, of maximum ice-affected (and open) water levels relative to a reliable benchmark.

5.2 Biophysical approach

This approach, described in NRCC (1989), mostly applies to breakup ice jams and to some freeze-up jams. It basically consists of investigating the elevation of ice scars on trees, vegetation trim lines, or debris lines. It seems that gradual floods caused by ice dams and hanging dams are not likely to affect the vegetation in a way that would contribute in delineating the ice-induced flood zone. The impact of recurrent, thick, overbank aufeis accumulation on riparian vegetation may enable the application of the biophysical approach, but this should be confirmed by field observations and interpretations.

5.3 Past flood extent approach

An easy to understand and defensible approach of flood delineation consists in mapping the extent of a past, ideally major, ice-induced flood events. Although often not associated with a specific frequency, it represents valuable information for decision makers, and together with a topographic map, the extent of slightly more intense events could be determined. This information depends on the spatial extent of flooding and (or) elevation of the maximum water level being measured or observed at multiple locations along a vulnerable stream reach. The elevation- or aerial extent- of ice-induced floods can be relatively easy to obtain when associated with gradual ice processes such as the formation of ice dams, hanging dams or aufeis (Grayson, 2010), especially along larger streams. For dynamic processes such as breakup ice jams, the flood elevation can also be estimated based on elevations of shear walls left on the banks. (Furthermore, information on the elevations of the water surface during the flood event is useful when calibrating hydraulic models). This is why resources (ground observers, drones, satellites) need to be deployed to document the spatial distribution of the water or ice elevation during and after a flood event, even though this has rarely been considered as an immediate priority of public security agencies in the past.

Figure 11 presents a flood map showing open water flood zones and the extent of a past ice-jam-flooding event along the Des Hurons River, a stream draining about 75 km² in Stoneham-Tewkesbury, north of Quebec City, Canada. While it is known that threatening ice processes can occur more frequently than open water floods at specific sites (e.g., ice jam upstream of the bend in Figure 11, where the flood plain is relatively wide), users of flood maps based on the past extent of ice-induced floods should be aware that a streamwise shift or a similar, but more intense, process can cause flooding beyond the outlined flood zone.

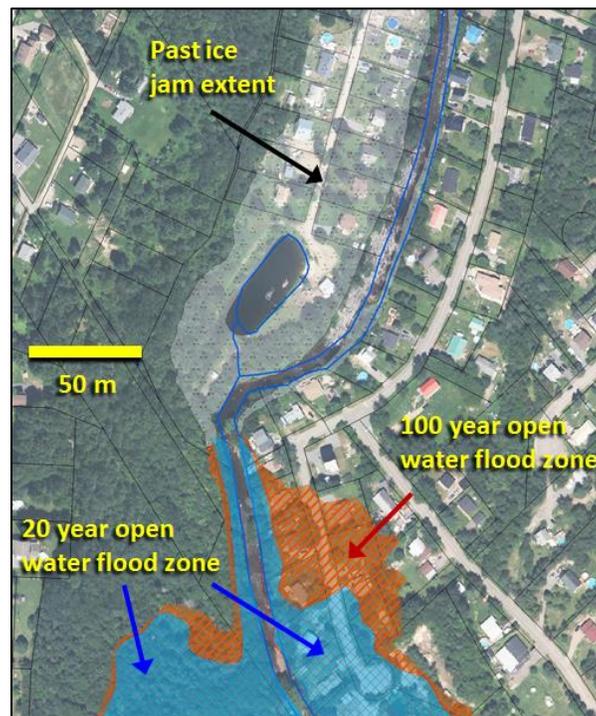


Figure 11. Extend of a past ice jam event in the Des-Huron River, Stoneham-Tewkesbury, QC, Canada, together with the 20 year and 100 year open water flood zone. Flow direction is downward.

5.4 Flood Envelope approach

This approach can be based on a combination of the previously described methods in order to extrapolate the extent of past and probable floods over long stream reaches. Although it is particularly applicable to areas of limited data or heterogeneous data sources, care must be considered to determine the flood envelope, especially along stream reaches with highly irregular topography. It is particularly difficult to apply in the case of floods generated by gradual ice processes in less developed or inhabited areas. In turn, for sub-arctic and arctic aufeis, thick ice sheets may remain in place during the summer and even year-round (e.g., Grayson, 2010), which provides enough time to map the flood envelop associated with aufeis development.

5.5 Hydrotechnical approach

The hydrotechnical approach involves some kind of simulation of ice processes that leads to high water levels. It is expected that the frequency of the annual maximum ice-induced water level should occur when the discharge rises while the stream channel is occupied by ice, but in specific reaches, the highest annual water level may occasionally occur at freeze-up (consolidated freeze-up jam) or during winter (hanging dams, aufeis), when the discharge may be close to its annual

minimum. This information must be known and documented in order to evaluate the frequency of high, ice-induced water levels.

In large rivers, the hydrotechnical approach has mostly been applied to evaluate the frequency of water levels caused by breakup ice jams, which typically control the annual maximum ice-influenced stage (Beltaos 2012, Gerard 1989, Tuthill et al 1996, White and Beltaos, 2008). Kovachis et al. (2017) refers to the equilibrium jam approximation as presented in Figure 10 for large channels (which evaluates the upper boundary of the red zone in Figure 9). In this case, the frequency of the water level (H) can be linked to the frequency of a unit discharge (q) in the presence of ice and to the frequency of ice jam occurrence. This can only be evaluated if: (1) a reliable discharge data set exists¹; and (2) ice jam dates are known. In the case of no (or limited) available discharge data and historical ice jam observations, these parameters could be indirectly determined using different models that would contribute in defining specific conditions and thresholds (e.g., the limits of the red zone in Figure 9). The development and operation of (conceptual, hydrodynamic, empiric or deterministic) river ice models are normally performed by experts who are familiar with the site of interest (e.g. Beltaos, 2008; Turcotte et al., 2015b; Tuthill et al. 1996, White, 2008). Note that the ice cover and ice jam mobilization discharges are among the most under examined and documented parameters associated with breakup, and yet they are probably among the most important to quantify.

For breakup ice jams in small streams (especially if narrow or steep), the general approach may be similar, but the equilibrium jam curve (Figure 10), or the ice jam computation in models such as HEC-RAS (Beltaos and Tang, 2013), may not apply (as discussed in Section 4.5). Assuming that some kind of ice jam frequency record exists, the corresponding water levels (H) could be:

- Computed with numerical models that are adaptable to different stream types: RIVICE (public domain), CRISSP and ICESIM (proprietary models).
- Estimated using common hydraulic equations or models. In turn, this could be based (1) on an evaluation of the discharge (Q) that leads to initial overbank flooding, considering a probable ice jam thickness and composite roughness, (2) on the assumption that any additional Q would not contribute to thickening of the ice jam, and (3) on the assumption that the ice would be confined in the channel. A comparable synthetic approach was proposed by Calkins (1983).

In most cases, some kind of empirical, often site-calibrated, function is needed to perform a representative hydrotechnical investigation and this requires considerable information gathering and data collection/ analyses, which makes the whole process expensive.

Although no synthetic equation, as presented in Figure 10 or in Calkins (1983), exists to apply the hydrotechnical approach in order to evaluate the frequency of water levels associated with other ice processes, the procedure remains similar:

- 1) Simulate the water levels associated with an ice blockage of specific mass of thickness, based on past measurements and discharge estimations¹. Note that for streams, the daily average discharge can significantly underestimate the peak flow and a representative peak factor (varying from 1.05 to more than 10) should be used if instantaneous discharge estimates are not available.

¹ The presence of ice effects on stream stage is clearly indicated in historical records, so as to warn users that associated flows are estimates and may not be as reliable as open water values. Missing data points are often associated with valuable extreme events.

- 2) Refer to the literature or develop a site-specific model that empirically links the process to other hydro-meteorological parameters for which the frequency can be determined (e.g., a discharge, a number of cumulated degree-days, a number of watts lost per square meters, etc.). This can only be accomplished if the physics of specific ice processes is mostly understood, which is not always the case (e.g., aufeis).
- 3) Create a relationship between the simulated water level and the frequency of the driving parameters.

Figure 11A presents a relationship linking the simulated amount of frazil slush, snow slush, and ice floes contained in a grounded hanging dam, based on an updated version of the model presented by Turcotte and Morse (2016) on the x axis, and a measured water level during 8 different winters on the y axis. The application of the same ice production-transport-deposition model to 40 years of hydrometeorological data yields the frequency of the ice accumulation in that specific reach (Figure 11B). The combination of both relationships yields the frequency of water levels associated with the formation of the hanging dam under low flow conditions (Figure 11C). This approach assumes that the ice produced by upstream reaches always accumulates following a similar pattern year after year, which can only be partially true (see the scatter in Figure 11A), and needs to be adapted to occasional mid-winter runoff events that produce higher water levels for a given amount of ice.

For thickened freeze-up jams, ice dams and aufeis, if no direct historical water level data (as presented in Figure 11A) is available, a reach specific blockage-frequency relationship could be developed, based on observations or numerical simulations, and it could be combined with discharge-frequency and ice decay-frequency empirical functions. Different ice experts would probably have different means to perform such an ice-induced flood frequency analysis.

In municipalities where a significant amount of damage has been reported on a relatively regular basis, independently of the origin of the flood, the hydrotechnical approach may be by-passed by a flood risk analysis and mapping (with reported costs associated with specific areas). Hybrid methods that consist of associating an average damage to a specific ice process and to simply simulate the occurrence or non-occurrence of that process can also represent a defensible flood risk determination option. The occurrence of specific ice processes may be confirmed by remote sensing approaches such as satellite images (e.g., Mermoz et al., 2009). The higher resolution of the new generation of satellites now enables the interpretation of the ice cover type along small rivers, but not for small streams where the use of ground observations, time-lapse cameras or drones will probably remain necessary.

The development of techniques to define the flood zone and the flood risk along streams depends on the availability of field data (as presented in 5.1). Therefore, there is a need to appropriately monitor river ice processes (Turcotte et al., 2017) on streams susceptible to high annual water levels predominantly caused by ice processes.

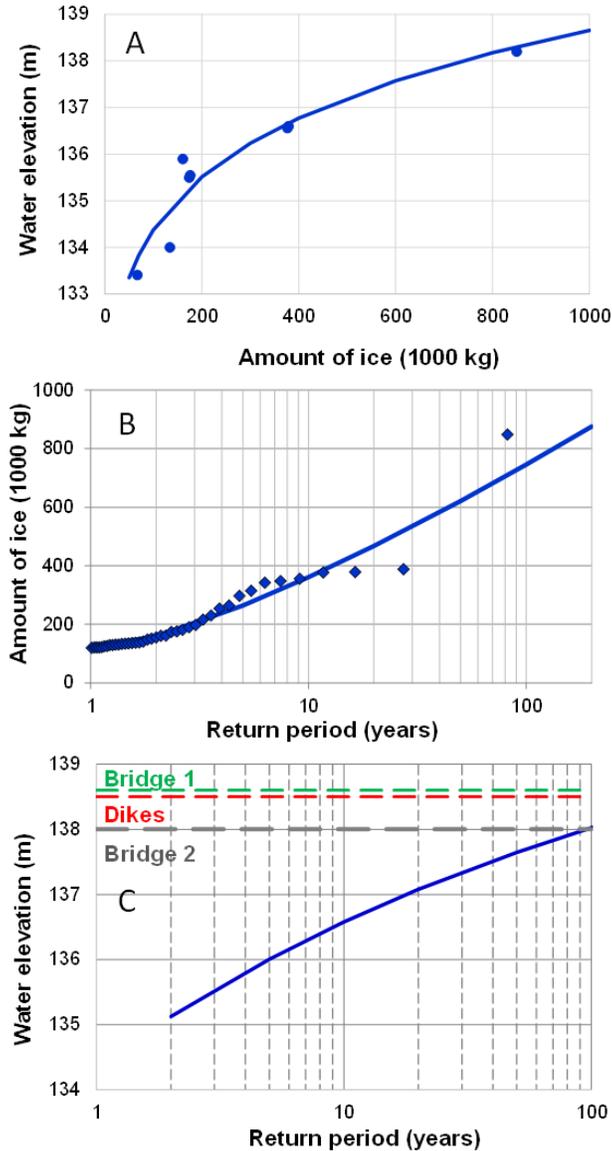


Figure 11. (A) interpolated relationship between measured water levels and simulated ice mass deposition, (B) relationship between the simulated ice mass deposition and the return period and (C) approximate relationship between the water level and the return period for a reach of the Ste. Anne River, QC, Canada (part of a project to evaluate and mitigate the risk of ice-induced floods).

5.6 Morphological approach

In the absence of local data or observations, one conservative technique to delineate the flood zone associated with different stream ice processes is to initially determine where damaging processes can occur using morphological or channel geometry indices (e.g., Bergeron et al., 2011; Turcotte and Morse, 2013) and then to perform a hydrodynamic simulation of these processes, based on realistic hydrological and ice conditions. Representative results could be obtained if a complete channel blockage up to the channel bank (for ice dams and aufeis) or a significant reduction in the channel evacuating capacity (for freeze-up and breakup jams that would eventually be lifted by the increasing discharge) is assumed, along with a potential flood plain snowpack.

This method has been used by Lier (2002) for the Karasjohka River in Norway. Potential flooding scenarios from ice jams were simulated using the hydrodynamic model MIKE-11. Three ice jam locations were tested, including two sites where no ice jam had been reported, but where conditions appeared conducive for ice jam formation, based on river topography. The findings from this study suggested that the ice-jam-flood frequency (and potential damage) was higher than the open water flood frequency.

6. Overview of Ice-induced Flood Mitigation along Streams

Engineers and community planners involved with land development should be aware that even small channels may become completely congested by ice, and that damaging ice-related flood events can result. The frequency and intensity of the ice-related flood hazard depends upon the causative ice processes, hydrometeorological conditions, and the hydraulic characteristics of the channel and flood plain. The damage of the ice-related flooding depends upon the amount of development in flood prone areas (i.e. exposed to the flood hazard), and the susceptibility of that development to flood damages.

Several measures can be used to reduce the risk of ice-related flooding and they can be classified from different perspectives (e.g., Turcotte and Morse, 2015a). More on the topic can be learned in Belore et al. (1990), Burrell (1995), Haechnel (1998), Tuthill (1995, 2013), and in White and Kay (1996). It is important to mention that mitigation measures successfully applied in large river environments may not be transferable to smaller channels, especially when they imply a modification of ice, hydraulic and hydrological conditions (e.g., Conventional [Abdelnour et al., 1998] vs. experimental ice booms [Lafleur et al., 2017]). A study done by Laval University for St. Raymond, QC, Canada (Turcotte and Morse, 2015a), is an example of an ice-induced flood risk assessment and mitigation study adapted to a large stream. More research is need to develop ice-induced flood guidelines specifically adapted to streams.

7. Concluding remarks

This paper has described different qualitative and quantitative ice processes that can cause flooding along streams and small rivers. Floods in streams and small rivers are caused by thermally-driven ice processes (mostly heat loss that generates ice formation leading to a gradual water level rise) by hydro-mechanical ice processes (rise in discharge causing a sudden ice movement and consequent rise in water level), or by a combination of both (e.g., freeze-up jams made of frazil [thermal process] and ice runs [mechanical process]). No certified and versatile approach exists to determine the limit of the ice-induced flood zone along streams and small rivers. A classic answer provided by river-ice experts is that every site and every ice jam is unique. That being said, the performance and adaptability of different approaches should be documented in order to provide a basis for future development of guidelines with respect to the flood hazard and risk mapping along ice-affected streams.

Municipalities that have suffered damage from ice-induced floods should consider documenting ice processes and their hydrologic impact in order to facilitate the development of adapted flood maps that are representative of ice-affected watercourses. In all cases, the damage associated with a specific water level should consider that various ice processes can generate even more damage for the same stage compared to open water floods (e.g., Humes and Dublin, 1988).

The effect of climate change on the ice regimes of streams may evolve more drastically than along larger rivers, generating more or less intense floods. Therefore, care must be taken when delineating the ice-induced flood zone along streams based on past data and observations.

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