

The concept of a maximum river ice breakup discharge - Investigating the hydromorphological context of applicability

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Executive summary

The main objective of this project is to explore an important question in river ice science and engineering: the concept of a maximum breakup discharge for a given river segment. Is it possible that, above a specific flow threshold, ice jams are always mobilized from a river reach, and can we link high and low thresholds to channel characteristics (i.e., morphology)?

To answer this question, hydrometric data from 2011 to 2025 for four distinct ice-affected rivers of Canada were analyzed (sites that are known for their ice-jam flood regime):

- Klondike River near Dawson, Yukon
- Hay River near Hay River, Northwest Territories
- Mackenzie River at Fort Simpson, Northwest Territories
- Chaudière River at Beauceville, Québec

Since this study involves a fine interpretation of historical flows during the breakup period, a technique that relies on the calculation and interpretation of the ice-induced backwater was used to reassess publicly available discharge time series.

The results suggest that an absolute maximum breakup discharge ($Q_{\max \text{ ice}}$) may not exist for the studied river segments, as an important parameter controlling breakup conditions can be identified in each case. Moreover, from a hydrological angle, $Q_{\max \text{ ice}}$ was equal to or significantly higher than the average annual maximum (or channel forming) flow (Q_2).

Inconclusive results do not mean that this research is over. It can be acceptable that, under extreme and unusual conditions, an ice jam could resist a driving force (i.e., flow) that is higher than what had been anticipated, and that $Q_{\max \text{ ice}}$, just like $Q_{\max}$ (maximum open water discharge in any given year), can be associated with a return period. However, the probability of a $Q_{\max \text{ ice}}$ to Q_2 ratio could still be linked to specific channel characteristics or morphologies.

This is what future research could clarify by analyzing a diversity of additional river sites that are more or less affected by ice jams (a combination of breakup ice jam flood regimes and open water flood regimes). A research framework is proposed at the end of the report to explain how this future research could be conducted.

Other important outcomes of this work also include:

- Several novel conceptual figures that describe breakup processes and the range of breakup flow probabilities. This includes material to inform the development of empirical breakup forecasting models.
- Figures illustrating why the assumption that post-ice season flows (often used in engineering projects) can be considered equivalent to maximum breakup flows or flows causing the maximum ice-affected stage may not be technically acceptable.
- The need to explore the roles of small and large javes (ice-jam release waves) and their associated unstable driving forces, can influence the identification of $Q_{\max \text{ ice}}$ (or $Q_{\max \text{ ice}}/Q_2$).
- The need to explore the interconnectivity between ice jams and the characteristics of cold-region river channels (i.e., how ice jams can influence channel morphology and vice versa).
- The need to account for river characteristics when projecting the future of ice jam floods.

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1. Context – the interest in maximum breakup flows

Floods are the costliest natural hazards in Canada and many other countries worldwide. In cold regions, ice jams account for a significant share of peak annual water levels (e.g., de Rham et al., 2025). This explains why ice jams must be included in flood mapping projects where they have been reported (NRCan, 2023) and why they represent an important hydrological phenomenon to forecast (e.g., Turcotte et al., 2025; White, 2008). Determining ice-affected water levels associated with specific return periods (e.g., 100 years), which informs the design of engineering structures, land development planning, and insurance rates, is not a simple task. This calculation usually relies on the hydrotechnical method (e.g., Kovachis et al., 2017), which involves the calibration and use of one or multiple ice-hydrodynamic models (e.g., HEC RAS, River1D, RIVICE, CRISSP; see review presented by Carson et al., 2011) to simulate floods caused by hypothetical ice jams. Those models depend on the measurement or estimation of several input parameters, among which the river discharge, or streamflow, is particularly influential: the higher the flow, the higher the water level, following a trend that is well above, and potentially steeper than, the local open water stage-discharge curve.

Most flood-mapping projects that define the parameters associated with high-intensity ice jams face a common issue: how high could the flow be in the presence of ice? This is an important question from three angles: 1. A good or useful hydrological record may not exist (i.e., no hydrometric station near the site of interest); 2. There may not be a major ice jam within the hydrological record (associated with a high flow and intense flooding), 3. The flow estimate in the presence of a historical ice jam may be significantly inaccurate (e.g., Turcotte and Rainville, 2021). Even if points 1 and 2 are checked, using rough flow estimates in an ice-hydrodynamic model may introduce significant uncertainty into the project results, and experts may remain unsure about how “bad” a “design ice jam” could look like. For instance, if one has observed a significant ice jam on a small river when the flow was approximately $100 \text{ m}^3/\text{s}$, is it possible that an ice jam at the same location would remain in place at a flow of $200 \text{ m}^3/\text{s}$ (and is this breakup flow associated with a return period of 20 years or 10,000 years)? Would the ice cover below the toe of the ice jam be strong enough to resist that additional hydrodynamic force? Adopting a conservative approach would mean answering “yes” to this question, but it could lead to the assumption that an impossible flood scenario is in fact possible, with consequences for property values and insurance costs.

A similar problem arises when using river ice breakup models to forecast ice jam floods. During a breakup event, if the model user (i.e., flood forecaster) is aware that an ice run is travelling towards a vulnerable site where intact ice is still present, the probability of flooding may depend on whether the ice run is intercepted by the local ice cover and forms an ice jam. Also, when an ice jam is in place and the flow is rising, which could represent an imminent or active emergency, would there be any insight into the maximum water level that could be reached in the next minutes or hours before this ice jam releases? Again, a conservative answer would be to assume that the ice jam would form or would remain in place as the flow rises. However, this assumption could yield one or multiple unwelcome and costly false alerts.

In both contexts, flood mapping and flood forecasting, the safety of the public must prevail, and a conservative assumption may be considered adequate. However, entirely ignoring the hypothetical concept of a maximum river ice breakup flow ($Q_{\text{max ice}}$) could undermine public trust and lead to

significant costs ultimately borne by taxpayers. Therefore, there is a need to shed light on this largely unexplored cold region's hydrological concept and to identify the morphological contexts in which ice jams may be mobilized at a flow that would be site-specific, and that may remain well below the flow that a river can produce following an intense rain-on-snow or early-season snowmelt event. This project aims to explore one of the most important gaps in the field of river ice breakup science and engineering, with the premise that the highest ice-affected stage occurs in the presence of a breakup ice jam when the gradually rising flow is just about to trigger a dynamic release.

Specific objectives of the project are to:

- Investigate the breakup regimes of different rivers in Canada, with special attention to flow in the presence of ice jams.
- Identify if and why the concept of a $Q_{\max \text{ ice}}$ could apply at those sites.
- Present a list of distinct cold-region river morphologies where this concept could apply or not.
- Propose a research framework to improve breakup flow estimates and to confirm the hypothetical existence of a $Q_{\max \text{ ice}}$ for distinct cold region channel morphologies.

2. Breakup flows – observation and modelling discrepancies

Let us first identify what appears to represent a scientific paradox in river ice: On one hand, most conceptual diagrams that present breakup ice jam water levels with their corresponding estimated flow do so in an area of possible backwater-discharge scenarios (or following an “ice-jam rating curve”) logically positioned above the open water rating curve. What matters here is that this area (or trend) clearly ends on the right at a flow that is generally lower than the maximum open water flow on record, as depicted or implied in Figures 2.1 to 2.4. On the other hand, several flood mapping projects that rely on the hydrotechnical approach to define a probability distribution for the discharge in the presence of an ice jam often consider that a maximum breakup discharge ($Q_{\max \text{ ice}}$) does not exist. This assumes that, on very rare occasions, a significant ice jam may sustain a discharge higher than previously documented or estimated, causing an unprecedented ice-jam flood.

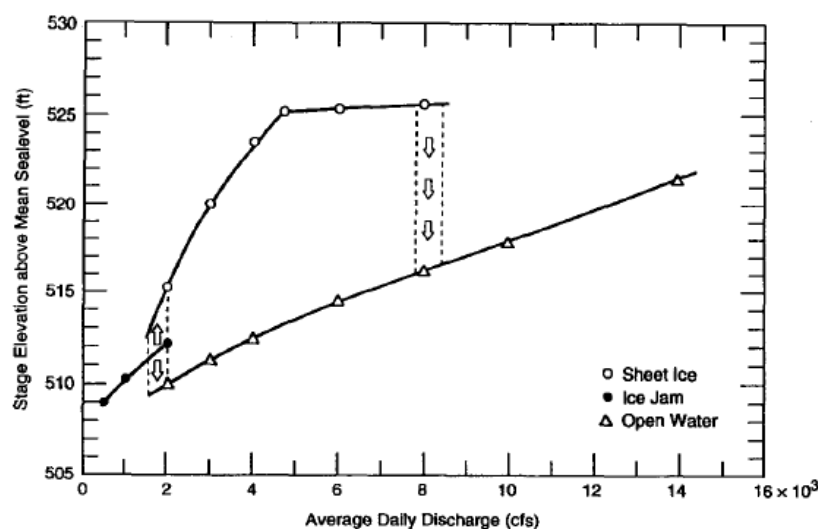


Figure 2.1. Reproduction of the conceptual diagram presented in Figure 1 in Tuthill et al. (1996). This paper was about the Winooski River in Montpelier, Vermont.

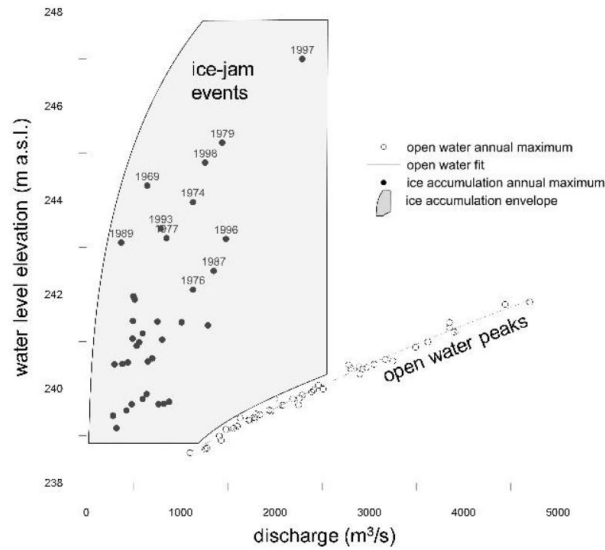


Figure 2.2. Copy of Figure 1 in Lindenschmidt and Rokaya (2019) presenting water level and flow data associated with historical ice jams on the Athabasca River at Fort McMurray, Alberta.

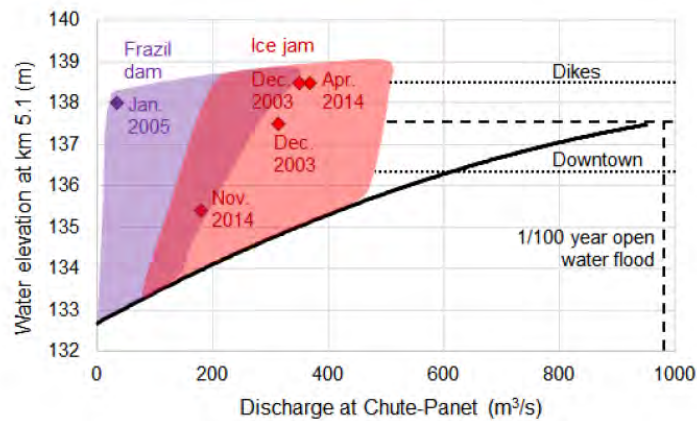


Figure 2.3. Reproduction of Figure 2 from Turcotte and Morse (2015). This paper was about the Ste-Anne River near Quebec City, Quebec.

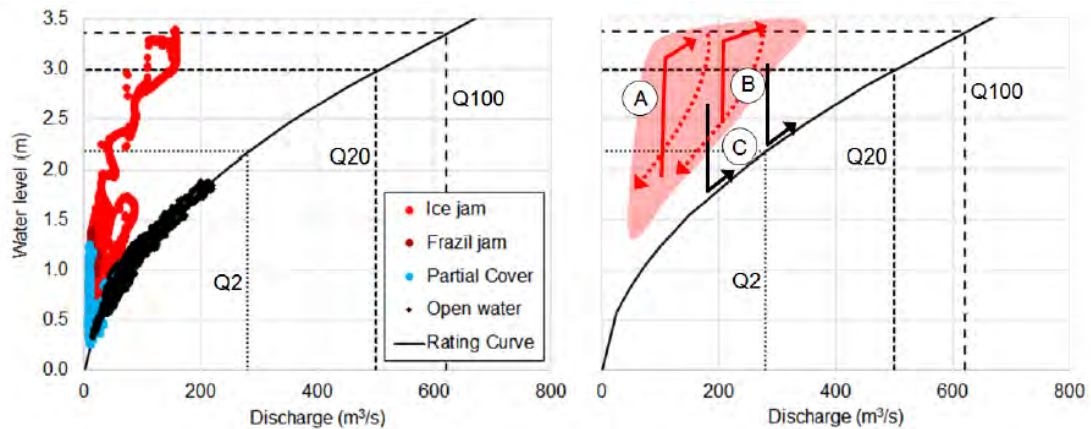


Figure 2.4. Copy of Figure 9. from Turcotte et al. (2017). This paper was also about the Ste-Anne River near Quebec City, but at a different site than the one associated with Figure 2.3 above.

Another contradiction in the field of river ice concerns the flow used in some hydrodynamic models to simulate intense, hypothetical ice jams. In the absence of accurate flow data during historical ice jam events (values are generally rough estimates and may contain errors; Turcotte and Rainville, 2022), modellers often prefer to adopt flows derived from the open water rating curve on the day following stationary ice clearance. This may be acceptable in some situations (e.g., where ice jams melt in place under stable flow conditions), but certainly not in all hydrological contexts, especially where and when the peak stage caused by an ice jam occurs several days or even weeks before the ice clearance is officialised by the agency responsible for the production of discharge data. This essentially means that the modeller tasked with simulating an ice jam uses a flow ($Q_{\text{post-breakup}}$) associated with a breakup-driving force that the jam (or downstream ice cover) may not be able to sustain ($Q_{\text{post-breakup}} > Q_{\text{max ice}}$).

The concept of a probable maximum flood in the presence of an ice jam, identified by Lindenschmidt and Rokaya (2019) as PMF_{ice} , is of great interest in the context of this study on $Q_{\text{max ice}}$. This study clearly indicates that “when a certain flow threshold is exceeded, ice jams become unstable and release, reverting to open water conditions”. Yet, their Figure 8, showing post-breakup flows as a function of the corresponding exceedance probability, does not appear to account for the existence of a maximum breakup discharge or an ice jam mobilization discharge in the Athabasca River near Fort McMurray. It is important to note that the RIVICE model used by Lindenschmidt and Rokaya (2019), as with HEC-RAS and other equivalent models, cannot “decide” whether and when an ice jam will release; the ice jam location and condition are imposed by the modeller.

Moreover, Lindenschmidt and Rokaya (2019) note that the PMF_{ice} identified in their study is associated with relatively “low discharge”. This somewhat defies the physically defensible concept of an ice jam rating curve presented in Figure 2.1. This may be the result of limitations inherent to the Monte Carlo approach, regardless of the number of simulation runs: 1. Input parameters are associated with a certain level of uncertainty, and their actual annual probability is commonly roughly estimated; 2. There might be some dependency between ice and hydrological parameters (e.g., volume of ice supply and breakup discharge), and by continuity, 3. Some input parameter combinations may be impossible. The Monte Carlo simulation results presented by Ladouceur (2022) for the Chaudière River near Quebec City illustrate that a methodology based on generated statistics can be highly inaccurate, regardless of the number of simulation runs, when the probability distributions of input parameters are not adjusted so the results match the stage frequency distribution derived from local measurements. Then, the adjustment of the probability of one parameter so that compiled results match local measurements could be considered as reasonable by some and as “forced” by others. Interestingly, when such an adjustment is possible, one may wonder why the distribution based on measured water levels is not used directly. For instance, Figure 2.5 presents an extrapolation of Figure 2 in Lindenschmidt and Rokaya (2019), with the simulated PMF_{ice} (250.43 m) advantageously corresponding to the approximate 1000-year event obtained from extending the Gumbel distribution (~250.5 m). This, however, does not provide information about the discharge corresponding to this uncommonly intense jamming event, nor can it confirm the existence of a $Q_{\text{max ice}}$ for the Athabasca River near Fort McMurray.

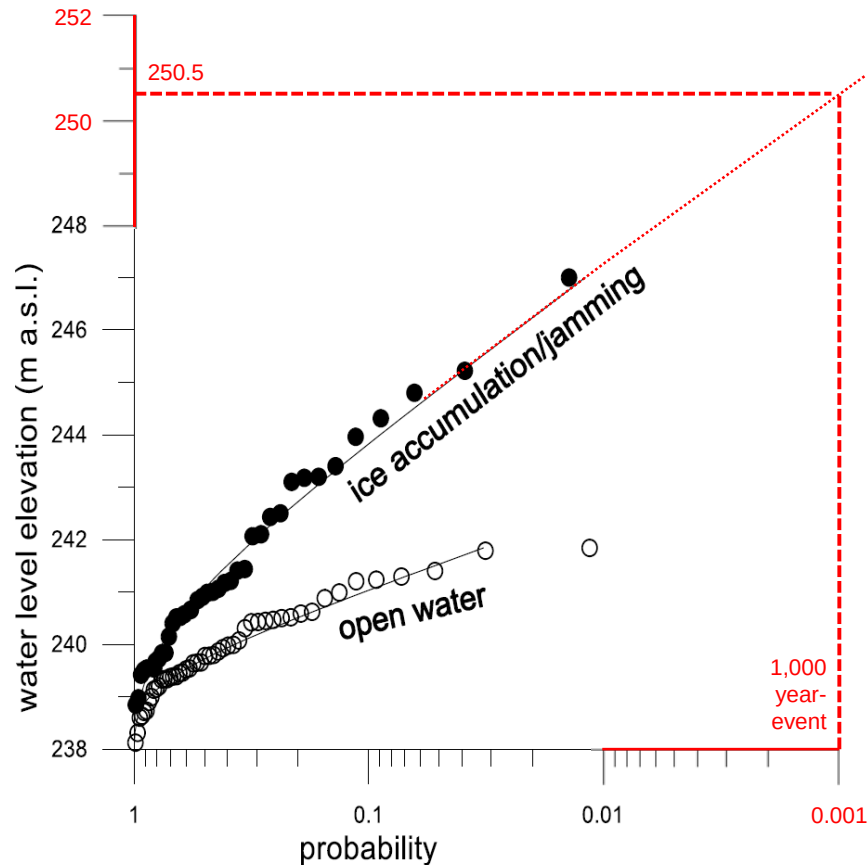


Figure 2.5. Extrapolation (in red) of Figure 2 presented in Lindenschmidt and Rokaya (2019) and showing a possible ice jam water level associated with an exceedance probability of 0.1%.

In a majority of cold-region river settings where ice jams have been observed, it is possible that maximum ice-induced stages are associated with low-intensity (thermal) breakup events or impeded ice runs that do not cause ice jams. The contribution of Healy and Kovachis (2023) is important for exploring the existence of $Q_{\max \text{ ice}}$, as it investigates the probability of ice-induced water levels across diverse breakup scenarios on the Peace River using the Monte Carlo approach, with mechanical breakup events causing ice jams statistically associated with high breakup flows.

Even the performance of complex ice-hydrodynamic models that simulate the forces acting on an ice cover downstream of an ice jam (e.g., CRISSP) can face limitations: it is difficult to calibrate a model if one does not know the forces that really occur at breakup, and users still need to impose some parameters that control these forces. Ultimately, it may be preferable to take a step back and investigate what rivers can reveal through direct historical measurements and observations, as proposed in the current study. The forces acting on historical ice jams may not be known, but the hydrometeorological conditions leading to breakup initiation, maximum ice-induced backwater, and ultimately ice jam mobilization at a specific site (or within a river segment) can be identified and analyzed.

Figure 2.6 returns the reader to the conceptual level of the ice jam backwater area positioned above an open water rating curve. The open water bankfull discharge, often considered a 2-year flow, or Q_2 , is shown as an important reference in this diagram. For instance, Q_2 on the Athabasca River at

Fort McMurray is roughly 2,700 m³/s, which probably not coincidentally corresponds with the right boundary of the ice jam area depicted in Figure 2.2 above. In Figure 2.6, this area is divided into three regions corresponding to increasing breakup resistance.

- I: Limited ice cover resistance (early winter period, mild winter conditions, significant ice cover degradation in the spring)
- II: Moderate ice cover resistance (low ice cover degradation, grounded ice jam)
- III: High downstream ice cover resistance (presence of a freeze-up jam, a mid-winter jam, or a hanging dam, water bypassing the ice jam toe, significant ice supply, etc.)

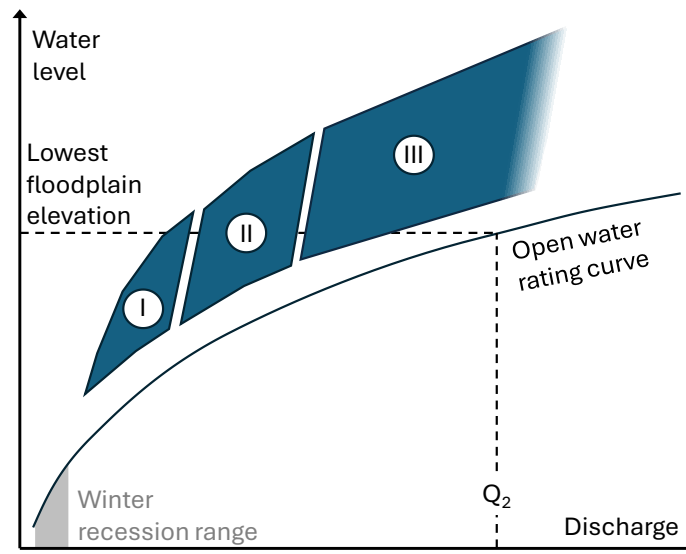


Figure 2.6. Conceptual diagram showing four distinct areas where peak ice jam stage may fall, depending on hydrological and morphological conditions.

The objective of this project is not to document the state of ice cover degradation at breakup across different rivers in Canada. Instead, the hypothesis is that the three regions shown in Figure 2.6 may also be associated with different river segment characteristics, classified by morphology (which implicitly includes hydraulic conditions and ice cover types; e.g., Turcotte and Morse, 2013). This means that, for a maximum ice cover resistance, the upper-right corner of each region would represent a reach-specific $Q_{\max \text{ ice}}$, based on channel morphology as well as the watershed's hydrological regime and climate.

For instance, to illustrate the intention of this study, it is safe to state that river segments composed of (single-channel) rapids correspond to Region I in Figure 2.6. This morphology is often associated with a steep gradient, fast-flowing water (thermal erosion), emerging rocks at low flow, and/or a small floodplain. These segments are generally the first to become ice-free as Q rises, and ice jams, if any, are often mobilized at a $Q_{\max \text{ ice}} \ll Q_2$. Many observation-based studies support this interpretation:

- Montmorency River upstream of Quebec City, Quebec (e.g., Turcotte and Morse, 2015)
- Ste-Anne River upstream of St-Raymond, Quebec (e.g., Turcotte and Morse, 2016)
- Chaudière River, multiple segments, Quebec (Pelchat, 2022, Zidan et al., 2025)
- Ross River near the community of Ross River, Yukon (Berti, pers.com., 2026)
- Hay River, multiple reaches, Northwest Territories (Kovachis, 2011)
- Athabasca River segments upstream of Fort McMurray, Alberta (Kovachis, pers.com., 2026)

Identifying cold-region fluvial contexts associated with Region III in Figure 2.6 is more challenging, and this is what this study explores.

Another important piece of background on the topic is that river ice breakup is generally snowmelt-driven in large rivers of Northwestern Canada, whereas it is often caused by rain-on-snow events in smaller watercourses of Eastern Canada. This means that ice cover resistance may remain significant in the East during spring (and mid-winter) breakup events, providing more opportunities to test the hypothetical existence of a $Q_{\max \text{ ice}}$ using historical data.

3. Methodology

This study is built around a relatively simple methodology divided into several steps:

1. Select hydrometric stations located on rivers presenting different hydrological regimes and morphological settings, and that have been regularly (often annually) affected by ice jams in recent years. Different hydrological regimes may include rain-driven or snowmelt-driven breakup, steep or low-gradient rivers, peaky or laminated (upstream lake) runoff hydrographs, etc.
2. Obtain basic hydrological and morphological information for the selected rivers.
3. Identify ice jam events for the selected stations using stage time series and satellite imagery. Gaps in stage time series are common at hydrometric stations in ice jam-affected reaches, further complicating the reconstruction of the chronology of (and hydrological conditions during) each breakup event.
4. Determine the representativeness of published discharge time series during the transition from intact ice cover to open water conditions by analyzing measured stage and back-calculated backwater time series (using the local rating curve). As stated by Turcotte and Rainville (2022) and, more detail by Turcotte (2021), the backwater (BW) and flow (Q_{est}) are linked by the following simple, linear equation:

$$BW = 1 - \frac{Q_{\text{est}}}{Q_{\text{rating}}}$$

Where Q_{rating} is the direct translation of the water level, or stage, into flow using the rating curve. It often occurs that whereas the breakup hydrograph (Q_{est}) presents a reasonable shape and trend, the backwater (BW) graph does not, indicating that Q_{est} is, in fact, inaccurate. For instance, BW should not increase significantly as the ice cover melts in place, and, in turn, BW cannot drop too quickly if the ice cover remains (as confirmed by satellite imagery).

5. Produce, when published Q (or associated BW) anomalies or inconsistencies are identified, revised Q_{est} using high-resolution stage data that allow for the identification of sharp changes associated with local ice movements (see example in Figure 3.1), considering the need for defendable Q_{est} and BW data series. Some historical Q corrections may be minor, whereas others may be significant. Flow data from other hydrometric stations can also be used.
6. Extract flow information from selected time series (originally published or reanalyzed) corresponding to the peak ice jam stage (generally during ice jam formation or ice jam release) for each station (see Figure 3.1).

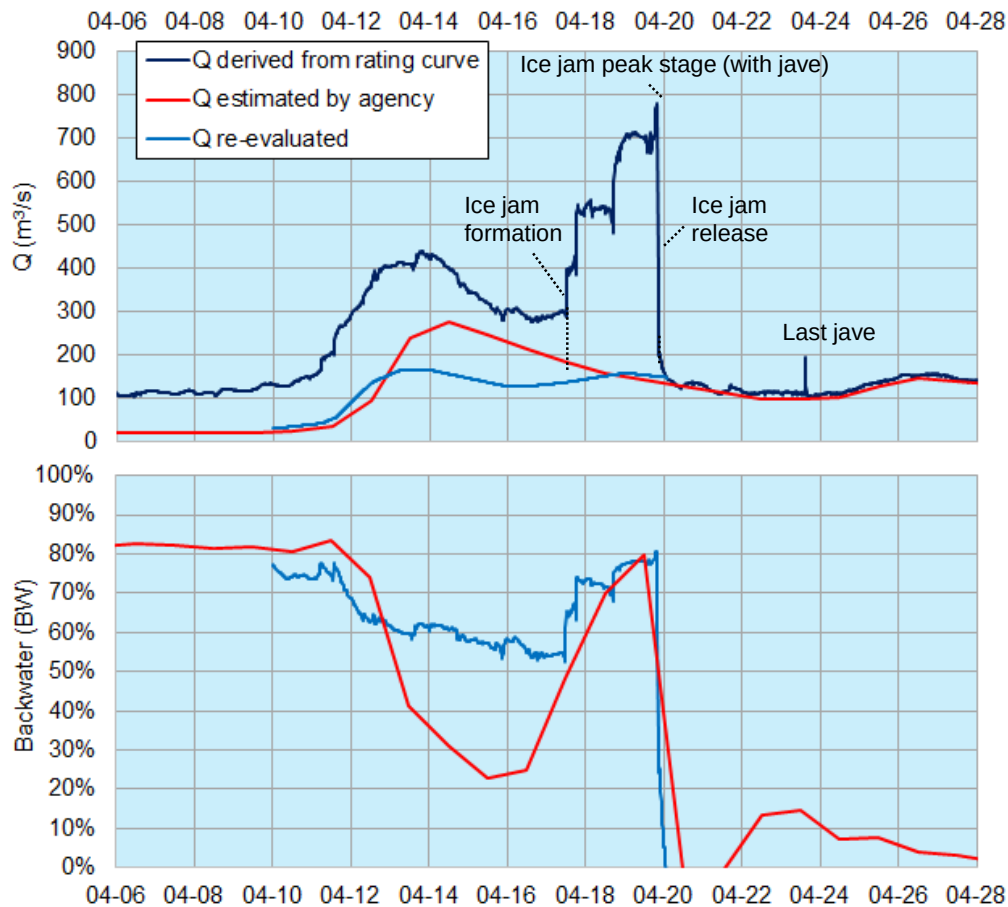


Figure 3.1. Example of breakup hydrograph including the rating curve-derived (or maximum physically possible) flow, published flow, and re-evaluated flow.

7. Compile (identify) hydrometeorological (morphological) data informing breakup resistance indicators for each breakup year (for each river setting). This may consist of islands, sharp bends, sharp enlargements, slope breaks (e.g., foot of rapids), etc.
8. Identify if the concept of a maximum breakup discharge ($Q_{\text{max ice}}$) seems to apply to each studied site and explain why.

Step 8 of this methodology could appear trivial if the results obtained from Step 6 indicated that spring breakup has never occurred above an identifiable estimated discharge at a specific station. However, this study takes this logic further down the river ice physics road: it explores channel characteristics that can control ice-resisting factors (at Step 7) beyond those associated with ice cover deterioration by heat or solar radiation, in an attempt to identify the influence of local or downstream morphology on extreme breakup events (taking place at high flow).

4. Analyses and results

This section presents the results of spring breakup analyses for four different sites located in Canada. All sites have been affected by significant ice jam flooding in recent years.

4.1. Klondike River – Yukon

The Klondike River (Tr'ondëk in Hän) flows westward on the Traditional territory of the Tr'ondëk Hwëch'in First Nation in Central Yukon. It drains a watershed of approximately 7,800 km² at the Water Survey of Canada (WSC) station 09EA003, located just outside of Dawson. Table 4.1 presents some hydrological and climatological statistics for the river, for context.

Table 4.1. Hydrological and climate information about the Klondike River near Dawson, Yukon.

Parameter	Approx. value
Average annual flow	75 m ³ /s
Average annual peak flow (Q ₂)	425 m ³ /s
Average annual minimal flow	12 m ³ /s
Average freeze-up date	Oct 25 – Nov 20
Average winter intensity	3400 °C-days
Average breakup date	April 25 – May 15
Date range of annual peak flow (mostly in open water conditions)	May 20 – June 10
Average annual air temperature	-3.5°C

The morphology of the river near station 09EA003 is significantly affected by historical placer mining activity (see linear tailings piles and randomly shaped tailings ponds in Figure 4.1). Along the first 1.5 kilometres upstream of the station, the channel is mostly laterally constricted by the north valley wall and by tailings piles with an average width of about 60 m. There is essentially no floodplain (only occasional lateral connections to tailings ponds), and the slope at low flow generally varies between 0.1% and 0.4% (average of 0.25%). The foot of a riffle upstream of the bridge is visible at low flow.



Figure 4.1. Aerial view (Google Earth) of the Klondike River just upstream of Dawson, Yukon.

The Klondike Highway bridge is located immediately downstream of station 09EA003. Below the bridge, the pool extends over some 750 m. Then the river widens significantly (from 45 to 160 m), and the channel splits into 3 steep (0.6%) subchannels. The station is therefore located at the upstream end of a pool, where a freeze-up jam tends to form annually, with occasional secondary consolidations (e.g., 2002, 2022). Breakup along the entire river is generally dynamic, with several small jams consolidating into fewer, larger jams, eventually leading to a stochastic release sequence. The breakup ice jam that forms near station 09EA003 is generally not the most resistant of the lower Klondike River, as reported by Turcotte et al. (2024): its resistance is often moderate, meaning that other ice jams remain in place under higher flow conditions.

The analysis of historical breakup events emphasized the period from 2011 to 2025. It first required a re-evaluation of ice-affected discharges (for 10 out of 15 years). Adjustments were generally in the order of $50 \text{ m}^3/\text{s}$ (25%) but could be as high as $250 \text{ m}^3/\text{s}$ (125%). Ice jams, staying in place for a couple of hours to almost a week, occurred during 13 of the 15 investigated spring breakup events. Peak stage at breakup was higher than during the subsequent open water freshet in 12 of 15 years, confirming the ice-dominated flood regime of the Klondike River at that location.

A 2021 study (Turcotte et al., 2021) used a discharge of $220 \text{ m}^3/\text{s}$ to simulate an approximate 200-year ice jam in the placer mining-affected area. This was not necessarily considered a proper $Q_{\text{max ice}}$ value but was identified as a high flow in the presence of an ice jam in the reach of station 09EA003. An initial estimate of $Q_{\text{max ice}}$ for the same reach was set at $250 \text{ m}^3/\text{s}$ by Turcotte et al. (2024), based on data from the 2023 breakup. Figure 4.2 presents the (re-evaluated) stage-discharge breakup data from 2011 to 2025, along with the local open water rating curve. $Q_{\text{max ice}}$ is now estimated at $270 \text{ m}^3/\text{s}$ (new 2023 evaluation). Note that, in most cases, the maximum breakup stage is reached when the ice jam releases, which explains why the red diamonds and black dots in Figure 4.2 are generally superposed. In two cases, the breakup peak stage is caused by a jave (ice jam-release wave from upstream) following local breakup. In 10 out of 15 years, the ice jam (or ice cover) mobilization takes place as the discharge is rising, whereas only two breakup events occur when the (reinterpreted) discharge (Q_{est}) is slightly declining. This is an important observation for that site, as it suggests that ice degradation is less likely to control jam mobilization than increasing hydrodynamic forces.

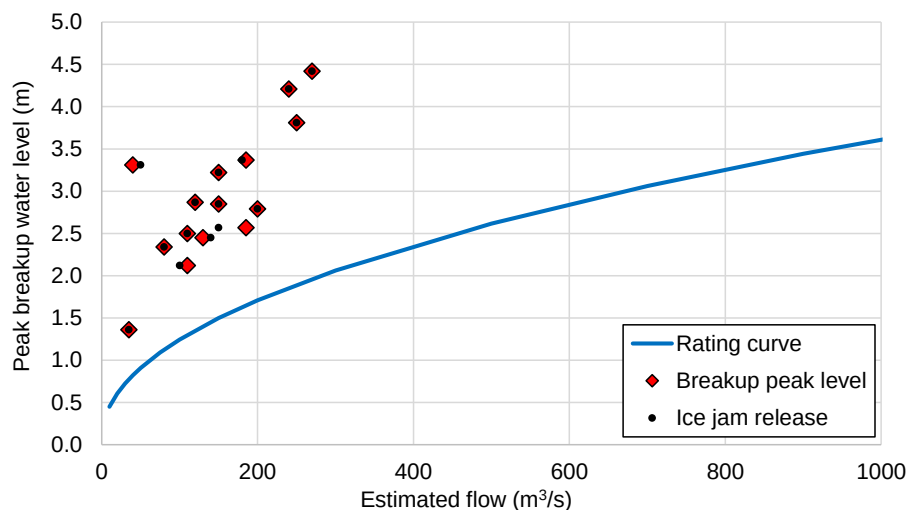


Figure 4.2. Open water rating curve (full historical range) as well as peak and ice-jam release stage-discharge data from 2011 to 2025 at station 09EA003.

Even after reassessment, uncertainty in the discharge associated with ice jam mobilization prevailed. This is mainly because of instabilities in the stage signal, which may be caused by local ice movements or by flow variations associated with upstream incoming ice jam release waves (or javes, lasting less than one hour). Therefore, minimum and maximum possible discharges were associated with each breakup event, as shown in Figure 4.3. This range was established by considering, just like Q_{est} described in Section 3, that both Q and BW should be defensible. The results suggest that, in some springs, the ice jam in the Klondike River was mobilized by a significant flow rise associated with an incoming jave and ice run. $Q_{max\ ice}$ in this dynamic context is now estimated at $450\ m^3/s$ (a 2013 data point).

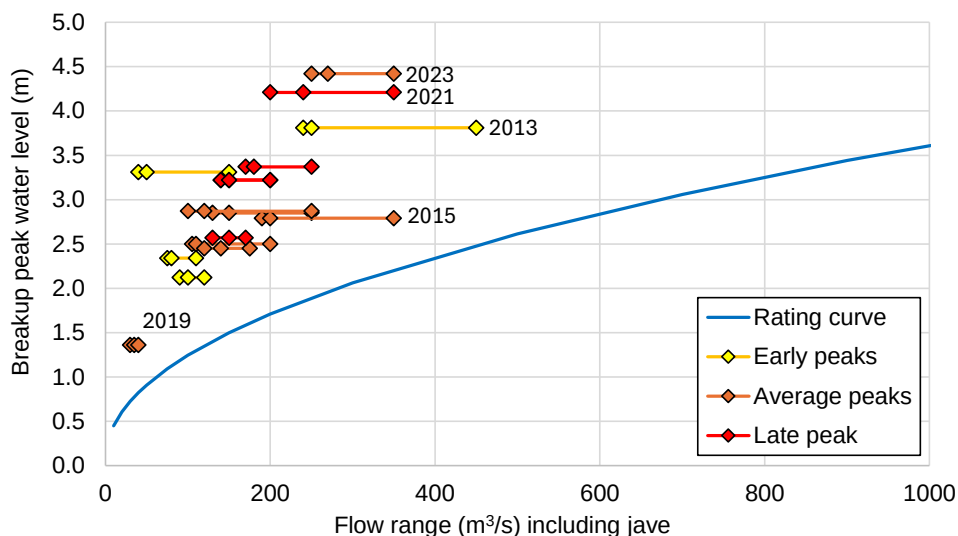


Figure 4.3. Possible range of flow associated with each ice jam (or ice cover) mobilization events from 2011 to 2025 at station 09EA003. Color range is linked to cumulated degree-days of thaw (early = few degree-days). The open water rating curve is presented for reference.

Figure 4.3 also identifies which breakup event occurred at low, average, or high cumulative degree-days of thaw (CDDT), based on statistics extracted from the local Environment and Climate Change Canada weather station. Results are not conclusive, with the most resistant ice jams associated with a wide range of CDDT. This means that weather conditions may not significantly influence breakup or a possible $Q_{max\ ice}$. Note, however, that the mobilization of the 2013 ice jam took place at only $50^{\circ}C$ -days of thaw compared with the average of $65^{\circ}C$ -days (range of 45 to 95° -days).

Results from Figure 4.3 also suggest that $Q_{max\ ice}$ could reach $500\ m^3/s$ in a dynamic breakup scenario, but under baseflow conditions, a $Q_{max\ ice}$ value of $300\ m^3/s$ may be more representative of this site. Three factors controlling breakup forces may limit this interpretation:

- **Snowmelt runoff limitation:** If the 2023 baseflow of $270\ m^3/s$ was essentially a runoff limitation in terms of snowmelt rate, it would not really inform on the maximum ice jam resistance. Fortunately, observations from that same year indicate that the flow can exceed $400\ m^3/s$ in the presence of ice. Moreover, rainfall during the breakup period (e.g., 1986) on the Klondike River could generate higher flows.
- **Warm water influence:** The presence of a long, open water segment upstream of an ice jam near station 09EA003, which is not uncommon, could allow the water to gain some heat. In turn, this could cause the jam to partially melt and release before Q further rises.

Observations from 2023 suggest that hypothetically warm water (e.g., 0.2°C) did not represent a factor impacting the stability of ice jams upstream of station 09EA003, with some jams remaining in place for a few days despite the presence of multi-kilometre open water segments upstream. That being said, this thermal effect cannot be ignored and may influence the results presented in Figure 4.3.

- **Downstream resistance:** As noted by Turcotte et al. (2024), with results replicated in Figure 4.4 for years 2011-2025, the freeze-up elevation appears to affect breakup resistance near station 09EA003. Moreover, the data suggest that, for a given freeze-up level (or freeze-up jam intensity), a high state of ice degradation (red diamonds) is generally associated with the least severe possible ice jam water levels. The breakup of 2021 (the highest red diamond in Figure 4.4), with its triple ice run interception (as interpreted from the stage signal), represents an anomaly in this case. Based on this analysis, one could conclude that if there were no limitations on the intensity of freeze-up, especially in a climate change context where rain-on-snow events may start to affect the central Yukon, then $Q_{\max \text{ ice}}$ may simply not exist for that site.

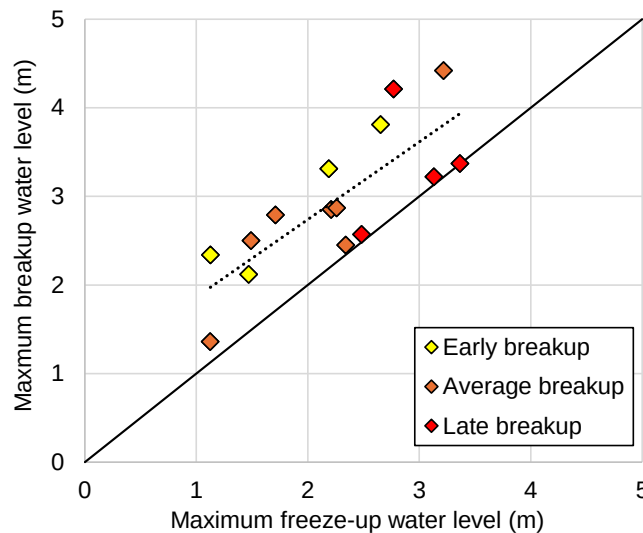


Figure 4.4. Maximum spring breakup level (mainly in the presence of ice jams) expressed as a function of maximum freeze-up level during the preceding fall at WSC station 09EA003 between 2011 and 2025. Colors represent the estimated ice degradation state based on cumulative degree-days of thaw.

In summary, does the concept of a $Q_{\max \text{ ice}}$ apply to the Klondike River near station 09EA003? The stage data in Figure 4.2 support this concept, with no data points above 300 m³/s. Even when looking at estimated data from 1970 to 2010, especially emphasizing the ice jams of 1986, 2003 and 2004, this threshold does not seem to be challenged. However, the analysis presented in Figure 4.4 suggests that a higher ice jam flow may be needed to mobilize a breakup ice jam that would have formed against a freeze-up jam, the intensity of which would be unprecedented. If the concept of a $Q_{\max \text{ ice}}$ existed at that location, it would likely be due to the single, narrow channel of the Klondike River combined with the absence of a floodplain that maximizes the downstream-oriented hydrodynamic force.

The results presented in this subsection are important in the context of flood mapping, flood forecasting, and even infrastructure design, because they suggest that a high-intensity (i.e., return period of 100 years) static breakup ice jam simulation near station 09EA003 should consider a discharge of 270 to 300 m³/s. Higher breakup flows could be expected for longer return periods. Figure 4.5 compares the flow at the time of peak ice jam stage, based in part on a re-evaluation of WSC daily-flow data using high-resolution stage data (refer to the example presented in Figure 3.1), expressed as a function of post-ice-clearance (post-B-date) flow published by the WSC for the Klondike River. The 50% average difference between the two data sets suggests that simulated ice jam water levels and flood extents derived from a frequency analysis that relies solely on published post-B-date flow data could be inaccurate and somewhat overly conservative.

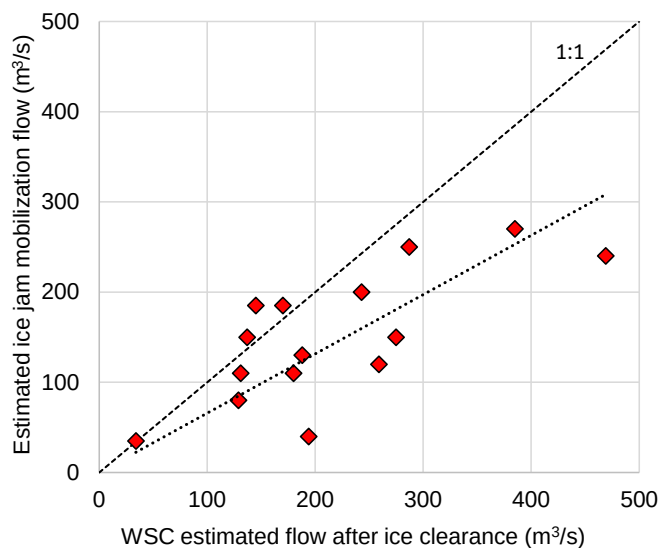


Figure 4.5. Re-evaluated flow (Q_{est}) at peak breakup stage compared to post ice clearance flow at WSC station 09EA003 for the period of 2011 to 2025. The interpolated trend shows a 50% positive departure on the WSC side.

4.2. Hay River – Northwest Territories

The Hay River (Kátł'odehche in South Slavey) flows northeastward across Northwestern Alberta and represents a major tributary of Great Slave Lake (Tu Nedhé [Big Lake] in Dëně Sųtıné, several other traditional names exist). Its lower reach in the Northwest Territories is home to the Kátł'odeeche First Nation. The river drains a watershed of 48,000 km² at the Water Survey of Canada (WSC) station 07OB001, located 10 km upstream of the current (new) Town of Hay River. The downtown area was relocated away from the lake in recent decades because the risk of ice jam flooding on Vale Island was too high.

Table 4.2 presents hydrological and climatological statistics for this river.

Table 4.2. Hydrological and climate information about the Hay River near the Town of Hay River, Northwest Territories.

Parameter	Approx. value
Average annual flow	110 m ³ /s
Average annual peak flow (Q ₂)	610 m ³ /s
Average annual minimal flow	3.7 m ³ /s
Average freeze-up date	Oct 20 – Nov 10
Average winter intensity	2700 °C-days
Average breakup date	April 25 – May 10
Date range of annual peak flow (often in open water conditions)	April 25 – June 15
Average annual air temperature	-2.1°C

Kovachis (2011) presented an exhaustive review of breakup processes and sequences on the Hay River, and also described the morphology of the river, including its downstream-most reaches (lower Reach 4 and Reach 5 in this thesis), which constitute the segment of interest for the current study. Hydrometric station 07OB001 is located 12.0 km upstream of Vale Island, which forms the river delta into Great Slave Lake. From the WSC station to the new downtown area (9 km), the river has a slope of about 0.02% and is composed of a few semi-confined meander bends that evolve into a fairly straight channel with high, vegetated banks. The channel appears very stable with few signs of lateral migration in recent decades. Short rapids are present in that reach (mainly visible at low flow), along with a few lateral and point gravel bars, but the river is generally slow-flowing and appears sediment supply-limited. The average width of the river is about 110 m near the WSC station, narrows down to 90 m at Pine Point Bridge, and then increases back to 160 m in front of the new town of Hay River. The downtown segment of the river is straight.

The delta is composed of two channels that play an important role in the sequence of river ice breakup scenarios (Kovachis, 2011). Their slope is approximately 0.01%; the main East channel presents a slightly milder slope because it is longer (6.5 km) than the West channel (5.5 km), which is narrower and often dry during the cold season. The delta includes a floodplain with oxbow lakes on the east side and vegetated sediment deposits that have been reshaped by the lake's waves over time. Based on elevation information, Great Slave Lake inundates most of the delta's channels at high water levels (e.g., above an elevation of 157.0 m), and flow velocities may be significantly reduced up to the new downtown area, especially during coincidental periods of low flow.

Breakup hydrographs at station 07OB001 were analyzed for the period from 2011 to 2025, which follows the period of 1894-2010 covered by Kovachis (2011). Only one of the 15 WSC breakup hydrographs was directly adopted, and the difference between the flow reinterpretation and the published WSC data mostly ranged from 5 m³/s (33% in 2024) to 120 m³/s (+70% in 2023), reaching 250 m³/s (-20% in 2022). Some of the largest corrections involved the inclusion of javes that accompanied the clearance of the ice cover in upstream reaches, whereas other corrections were negative, a decision informed by the analysis of the backwater time series derived from station 07OB001's rating curve. Satellite imagery from Sentinel 1 and Sentinel 2 was also used to obtain information about ice conditions upstream, at, and downstream of the station. Data from another station, 07OB008, located about 150 km upstream near the Alberta-Northwest Territories border, were also used to support the reconstruction of breakup hydrographs. The river segment where that

upstream station is located is usually ice-free by the time breakup occurs near the Town of Hay River, and the shape of the snowmelt hydrograph can occasionally be inferred from that open water dataset. Annual peak water levels at station 07OB001 occurred in the presence of stationary ice (mainly ice jams) in 12 of the 15 years included in this study, thereby confirming the dominance of ice jams as a flooding process near the Town of Hay River (also see Turcotte, 2021).

Figure 4.6 presents the measured peak breakup stage and the corresponding reinterpreted discharge at station 07OB001 from 2011 to 2025. The local open water rating curve is also presented. It is interesting to note that the maximum flow at that station since 1964 occurred just after the 2022 ice jam release (1740 m³/s). The breakup hydrograph and peak freshet flow were driven by a significant rain event in the headwaters (about 40 mm of rain measured at High Level, Alberta, on May 6-7) and were the highest on record, by far. It is possible that an ice jam-release wave (jave) from upstream, occurring when the massive 2022 ice jam was in place, could have carried a discharge higher than the subsequent freshet peak.

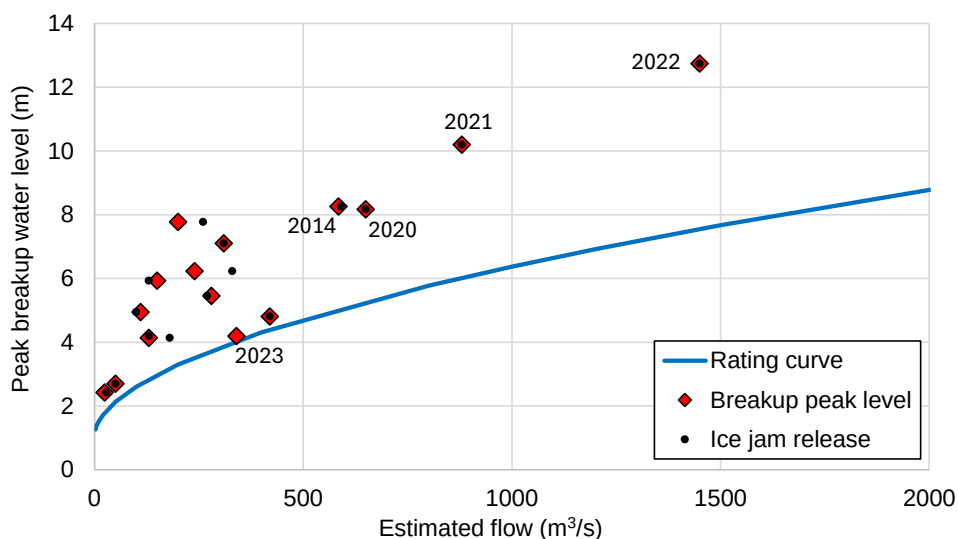


Figure 4.6. Open water rating curve (full historical range) as well as peak and ice-jam release stage-discharge data from 2011 to 2025 at station 07OB001.

Here is a summary of the interpretation of the stage time series and satellite imagery for that period:

- In 10/15 years, the ice jam melted in the delta as the flow was either rising or stable (what Kovachis, 2011, defines as a “melt out”).
- In 4/15 years, the ice jam seemed to release dynamically, with the flow either rising or stable. Figure 4.7 presents a Sentinel 2 image of the 2022 ice run coinciding with the stage drop at station 07OB001. This ice run is flowing into Great Slave Lake through the small West Channel.
- In 2023, a significant jave (periodic flow rise of 200 m³/s, more than 100%) occurred with little ice left in the delta. The presence of an ice-induced backwater during this event is uncertain.

This means that over the last 15 years, ice jams that formed at or downstream of the station melted in place about 75% of the time. This would tend to make the identification of a $Q_{\max \text{ ice}}$ more difficult. However, of the 4 most significant ice jams (maximum stage above 8 m at station 07OB001) over the last 15 years, two (2020, 2022) appear to have released dynamically. The data point of 2022 is especially important for investigating the possibility of a $Q_{\max \text{ ice}}$ on lower Hay River.



Figure 4.7. Sentinel 2 image (Copernicus, 2026) of the Hay River delta on May 12 showing an ice run flowing in the West Channel into Great Slave Lake. The water level at station 07OB001, located 12 km upstream, dropped significantly through that day.

Just like what was presented in the previous subsection for the Klondike River, the uncertainty of the published flows combined with the occurrence of small javes in the presence of an ice jam at or downstream of station 07OB001 imposes a certain level of cautiousness when investigating breakup flows. A range of estimated possible flows associated with the melt or release of each (minor or major) ice jam is presented in Figure 4.8. Also shown is an ice cover deterioration indicator (cumulative degree-days of thaw) for each jam at the time of peak breakup stage. When excluding data from 2021 and 2022, the message is that a low CDDT value (Early peaks; less than 25°C-days for the Hay River at the Town of Hay River) maximizes the chances of high breakup water levels (or an intense ice jam). However, ice jams from 2021 and 2022 are significantly blurring this signal. A similar analysis had been performed by Kovachis (2011) for the period of 1951 to 2010 (with gaps), with results (Figures 3.17-3.19 in the thesis) that were essentially the same: unclear. Importantly, the range of CDDT at the peak breakup stage for the lower Hay River is relatively narrow, with most events happening between 15 and 60°C-days (this study and Kovachis, 2011). Clearly, under such a narrow range, other factors are likely to explain the intensity of breakup.

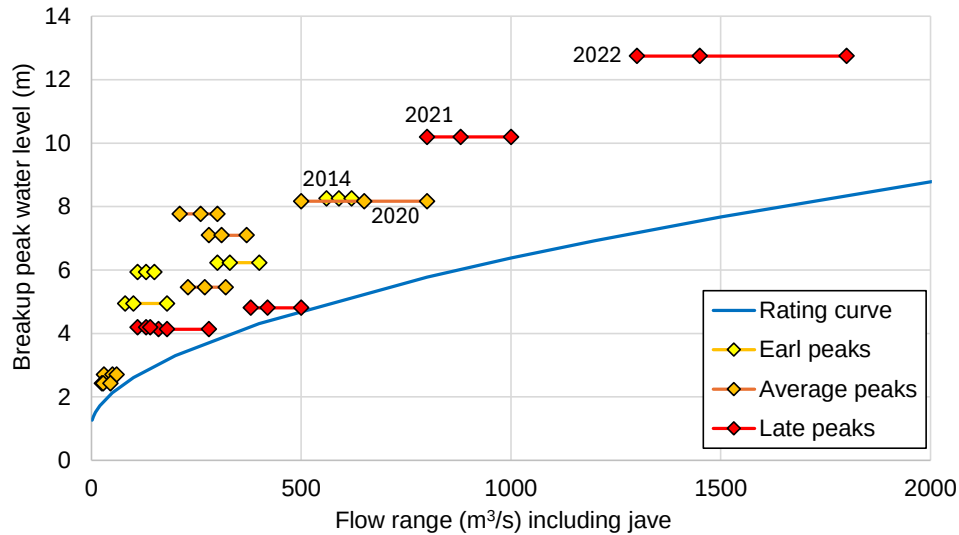


Figure 4.8. Possible range of flow associated with each ice jam (or ice cover) melt and mobilization events from 2011 to 2025 at station 07OB001. Color range is linked to cumulated degree-days of thaw (early < 25 degree-days; late > 45). The open water rating curve is presented for reference.

Figure 4.8 suggests that an absolute, reach-specific $Q_{\max \text{ ice}}$ does not exist for the lower Hay River simply because the ice jam that flooded the Town of Hay River in 2022 resisted a flow that also corresponds to an all-season historical record. However, it is of interest to briefly investigate whether this hydrological threshold could exist under certain conditions.

The author of this report was interested in the record-high water levels of Great Slave Lake from 2020 to 2022, especially during breakup events. Figure 4.9 presents the maximum breakup water level at station 07OB001 expressed as a function of the water level in Great Slave Lake on the same day (WSC station 07OB002). This was not of interest from a groundwater table (or watershed-scale hydrological) perspective, but rather from a river-gradient angle related to breakup-resisting forces. Interestingly, the results reveal that in the presence of record-high (2021, 2022) and extreme-low (2024) lake levels, a clear positive trend emerges. The colour of each diamond, related to CDDT just like in Figure 4.8, suggests that the degradation of the ice cover that would have occurred under warm air temperatures has essentially no influence compared with the downstream resistance that a high lake level seems to impose on breakup intensity.

To confirm this trend using a longer record, a similar analysis was applied to the year range covered by the Kovachis (2011) thesis, with results presented in Figure 4.10. When adding 38 years of data to what is presented in Figure 4.9 (1959-1970 and 1984-2010, limited by WSC operation years), a striking relationship appears: 86% of the significant historical floods at the town of Hay River occurred when the lake level was above 156.8 m, and 94% of all floods took place when this level was higher than 156.6 m. In turn, the probability of flooding was lower than 10% when the lake level was below the same threshold of 156.6 m. These thresholds generally agree with what Figure 4.9 reveals.

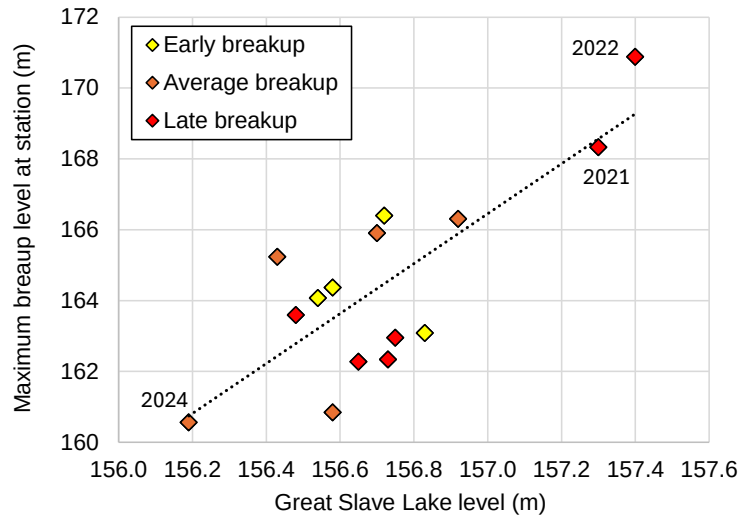


Figure 4.9. Maximum breakup water level at station 07OB011 on the Hay River expressed against Great Slave Lake water level on the same day (data from 2011 to 2025).

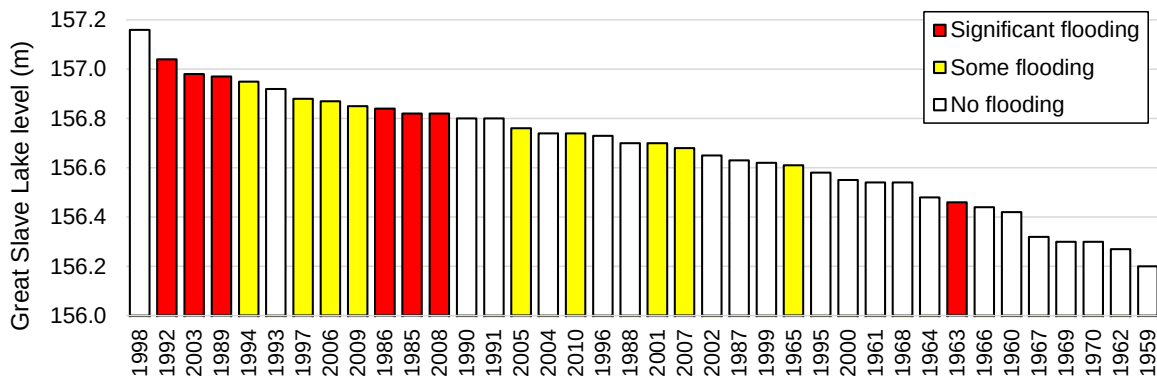


Figure 4.10. Great Slave Lake water levels at station (07OB002) with related qualitative flood assessments at the Town of Hay River as presented by Kovachis (2011).

What Figures 4.9 and 4.10 indicate is that a $Q_{\max \text{ ice}}$ may exist at station 07OB001, but it may vary as a function of the Great Slave Lake water level at the time of breakup (this excludes the water or ice level spikes generated by the breakup of the Hay River). Since the water level in Great Slave Lake represents a downstream boundary condition that could be compared to a time-varying, downstream channel geometric (or morphologic) characteristic, the hypothetical concept of a $Q_{\max \text{ ice}} = f(\text{Lake level})$ cannot be discarded. This is beyond the scope of this study, but readers will note that a simple seasonal ice jam flood forecast model can easily be derived from the data presented in Figure 4.10.

It would be of great interest to swap the maximum ice-induced stage at station 07OB001, as presented in Figures 4.6, 4.8, and 4.9, with the maximum water level measured at the community of Hay River or at another downstream reference location (the author of this report did not attempt to access such data). It would also be of interest to obtain such data points for the period covered by Kovachis (2011), as these modifications and additions could generate a noticeable reshuffling of the data presented in each figure. However, it would be unlikely to change the preceding interpretations.

In addition to what is presented in Figures 4.9 and 4.10, the most relevant breakup intensity proxy at the Town of Hay River, as explored by Kovachis (2011), is the coldness of the pre-breakup period (from April 1st until breakup in late-April or early May). This represents a breakup resistance factor. Other factors that have not been explored here nor by Kovachis (2011) and that could become part of a breakup intensity forecast model for the Town of Hay River include:

- Late-winter snowpack (to determine whether the breakup drive has the potential to be significant)
- Freeze-up or mid-winter discharge (that may limit the volume of ice in the river at the onset of breakup, which represents a breakup driving indicator)

Both parameters can be measured or evaluated well before the breakup period, just like the Great Slave Lake level. Then, the discharge of the Hay River during the breakup period could be estimated, and an envelope of possible water levels above the open water rating curve could be extracted from Figure 4.8.

4.3. Mackenzie River at Fort Simpson

The community of Fort Simpson in the Northwest Territories is located on the shore of the Mackenzie River (Dehcho in Slavey) immediately downstream of the confluence with the Liard River (Nácháh Dehé). This is the home of the Dehcho Dene people of the Liidlii Kue First Nation (meaning “The Place Where the Rivers Come Together”). Both rivers flow in a generally northward direction before reaching Fort Simpson. However, one cannot presume that this single aspect influences their spring hydrograph as is the case for the Hay River (Section 4.2):

- The discharge in the upper Mackenzie River (upstream of the Liard River confluence) is largely governed by the water level in Great Slave Lake, with significant influence from the regulated Peace River (Gibson et al., 2006), and by the breakup dynamics of its outlet (juxtaposed types of ice jams are frequent along a 300 km river section that include many channels widening and narrowing). This watershed extends over roughly 1,000,000 km² and generally supplies 45% to 65% of the breakup flow at Fort Simpson.
- The Liard River’s watershed is located west of Fort Simpson, including some high-elevation terrain where snowmelt is delayed. However, tributaries of the Liard River, including the Fort Nelson River and the Kechika River, drain the southern edge of the watershed and likely contribute significantly to the early-season snowmelt hydrograph at the Liard River outlet. This hydrograph is sharper than that of the upper Mackenzie. The Liard River drains a watershed of 275,000 km² and supplies 35% to 55% of the breakup flow at Fort Simpson.

Table 4.3 presents hydrological and climatological data of the Mackenzie River at Fort Simpson.

The lower reach of the Liard River at Fort Simpson has an average slope of about 0.025%, a value that generally declines downstream before it meets the Mackenzie River (NHC, 2025). The width of the channel varies between 550 m and 800 m near the ferry crossing before gradually increasing to a total width of more than 2,000 m at the confluence, including a set of small islands (called Truesdell Islands) on the side of which ice jams often form (Figure 4.11). The reach presents a single bend below the ferry crossing, and there is essentially no floodplain to accommodate overflow during high water level events.

Table 4.3. Hydrological and climatological information about the Mackenzie River at Fort Simpson, Northwest Territories.

Parameter	Approx. value
Average annual flow	6,850 m ³ /s
Average annual peak flow (Q ₂)	17,800 m ³ /s
Average annual minimal flow	2150 m ³ /s
Average freeze-up date	Nov 20 – Dec 10
Average winter intensity	3000°C-days
Average breakup date	April 25 – May 15
Date range of annual peak flow (often in open water conditions)	May 20 – July 15
Average annual air temperature	-2.4°C

The reach of the Mackenzie River upstream of the Liard confluence is relatively flat (~0.008%), straight, and narrow (1,000 m), with a few islands. The river steepens noticeably (~0.016%) near and below the confluence (NHC, 2025). Below Fort Simpson (essentially located on an island with a secondary channel blocked by a berm), the Mackenzie River widens to about 1,500 m on average and presents tight and short bends (not proper meanders; the bends are largely imposed by the surrounding geology and topography, Figure 4.11). Just as for the Liard River, the wetted width of the Mackenzie River does not increase significantly at high flow or in the presence of significant ice jams.

Discharge (and stage) data during the breakup period at station 10GC001 (Mackenzie River at Fort Simpson) contained several gaps, which is probably a consequence of dynamic ice processes affecting the station integrity. Breakup hydrographs were reconstructed (or re-evaluated, when the record was continuous) from 2011 to 2025 using other Water Survey of Canada hydrometric stations, mainly 10FB006 (Mackenzie River at Strong Point, located 30 km upstream of Fort Simpson) and 10ED002 (Liard River near the Mouth). In several cases, the estimated flow (Q_{est}) from stations 10FB006 and 10ED002 needed to be reassessed (using the approach briefly described in Section 3) to define the combined hydrograph at station 10GC001. Flow corrections at Fort Simpson could be as high as 5,000 m³/s (~50%) but were generally under 2,000 m³/s (or 25%). Major javes (sharp stage increases and drops generally originating from the Liard River breakup) or fast rises in snowmelt runoff (smoother, prolonged stage increases) are generally the most difficult to identify (or distinguish from one another) in a breakup stage record. This confirms that estimating river flow when ice conditions change dynamically (by the hour) is challenging, but in the end, the retained stage signal interpretation must be based on all relevant data (e.g., satellite imagery), and the associated backwater graph must make sense as well. Ice jams caused the annual peak water level at Fort Simpson for 13 or 14 of 15 years (only 2023 remains uncertain), which again confirms the breakup-dominated flooding process identified and simulated by NHC (2025) for the community.

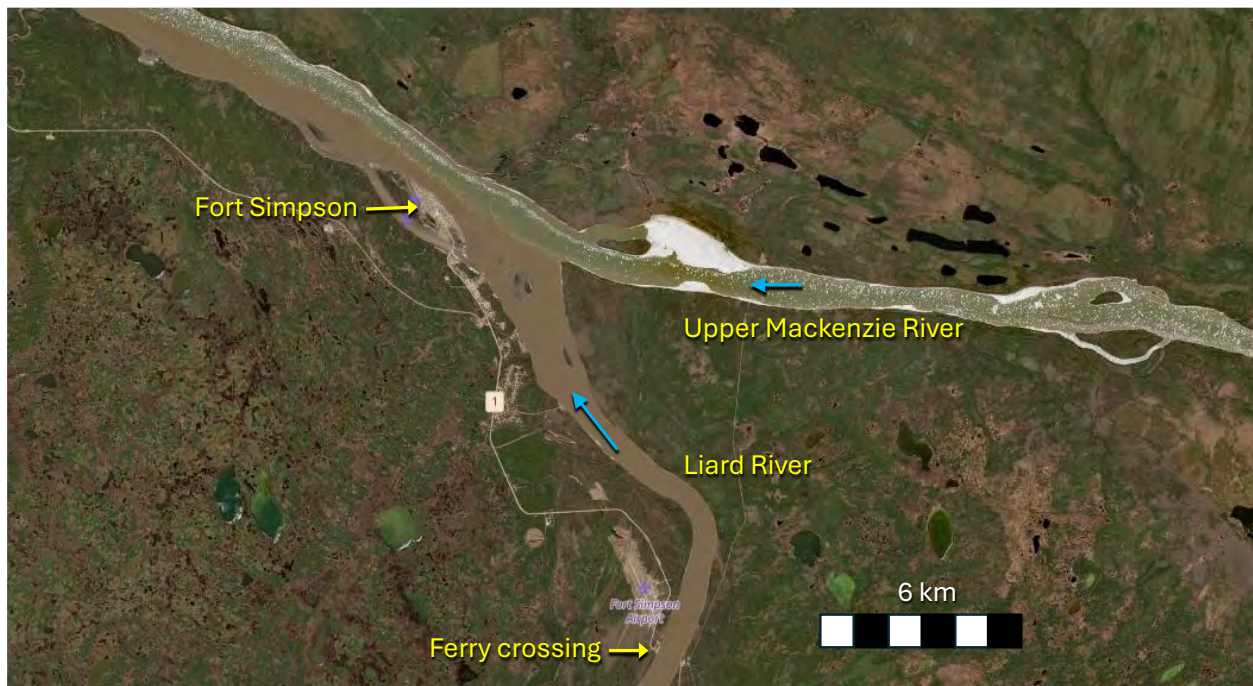


Figure 4.11. Sentinel 2 image (Copernicus, 2026) of the Liard-Mackenzie confluence on May 21, 2021 (high flow).

Breakup on the (steeper and hydrologically peakier) Liard River is generally dynamic, and the resulting ice run often accumulates against the Mackenzie River ice cover or in front of Fort Simpson. In some years, although the ice jam toe is located upstream of the community, backwater is detected at station 10GC001, likely due to ice floes settling under the ice cover. For the studied period of 2011-2025, breakup in front of Fort Simpson was triggered by the Liard River at least on 11 springs. This implies that the role of the Mackenzie River upstream of Fort Simpson is generally more thermal, with a strong hydrological influence from Great Slave Lake, as stated above. The presence of a freeze-up consolidation just upstream of the Liard River confluence may also contribute to this dynamic.

Measured (or reported) peak breakup water levels are expressed in Figure 4.12 as a function of their corresponding estimated discharge at station 10GC001 for the studied period. For context, the highest estimated flow on record since 1970 is about 26,000 m³/s. Unsurprisingly, the major ice jam of 2021, which took place when the baseflow (without ice-induced flow fluctuations) was 16,000 to 18,000 m³/s, represents a rare event at Fort Simpson. It is possible that a jave or a sudden increase in runoff from the Liard River could have caused the flow to rise to 20,000 m³/s while the jam was still in place (stations were down at that time). The jave of May 18, 2021, which probably came from a jam in the upper Mackenzie River, occurred after the release of the jam at Fort Simpson, causing the flow to rise rather sharply to about 24,000 m³/s. The disposition of black dots in Figure 4.12 generally indicates that the peak breakup stage coincided with the same driving force that triggered the release of the local ice jam (or ice cover, in the case of impeded ice runs). In 5/15 years, the discharge had further increased after ice jam mobilization and ice clearance.

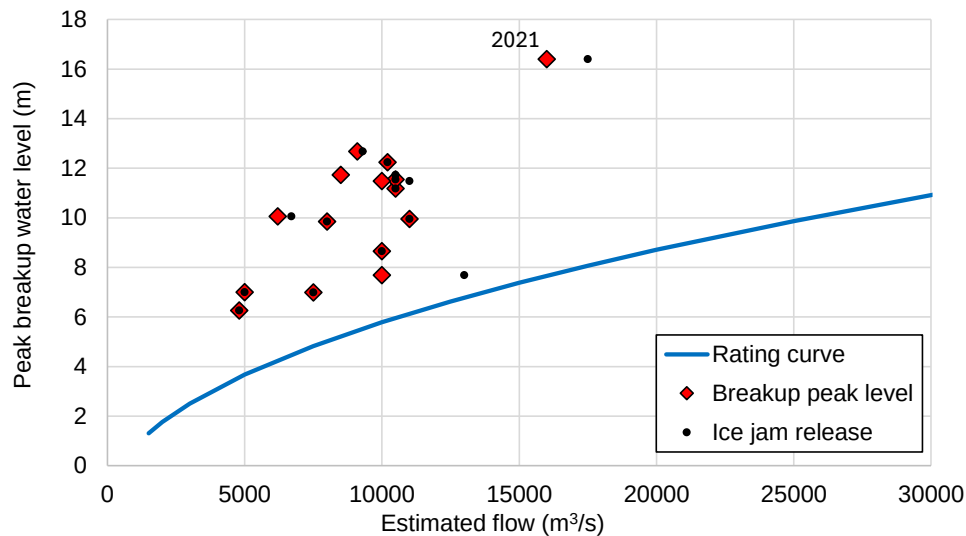


Figure 4.12. Open water rating curve (1970-2025 range) as well as peak and ice-jam release stage-discharge data from 2011 to 2025 at station 10GC001.

Figure 4.13 presents information similar to that shown in Figure 4.12, but with error bars reflecting approximate flow uncertainties. These bars also account for the presence of javes when an ice jam was in place or when station 10GC001 was unavailable. The approximate CDDT (cumulative degree-days of thaw), an assumed indicator of ice cover degradation, is also shown for each breakup event at the time of peak stage, using a colour indicator. Since 2011, the peak breakup stage has occurred between 50 and 125°C-days of thaw at Fort Simpson. The 2025 peak breakup stage is uncertain; therefore, a vertical error bar is also associated with this single data point. A result that is similar to what was presented for the Hay River is visible in Figure 4.13: The state of ice deterioration, estimated through an air temperature indicator (going from yellow to red), seems to be linked with backwater (departure from the open water rating curve) at a similar discharge. However, the 2021 event blurs this signal: it occurred at 137°C-days on May 12, which means that the ice cover should have been fairly deteriorated. In other words, one would have expected the 2021 ice jam to be mobilized at a lower driving force, or flow. On satellite imagery, the ice cover at Fort Simpson appears undeteriorated in early May, but air temperatures remain consistently above freezing starting on May 4. The jam forms 3 days later, and the stage continues to rise in the following days. So, what could have caused the jam to be more resistant than anticipated?

The role that the freeze-up stage may play during the following breakup season was investigated. Figure 4.14 presents data from 15 years. The positive relationship is similar to that shown in Figure 4.4 for the freeze-up stage of the much smaller and steeper Klondike River and in Figure 4.9 for the downstream lake level at the delta of the smaller Hay River. The length of the data set represents a limitation; thus, it is not surprising to see a relatively compact cloud of points for more common freeze-up and breakup conditions. Notably, early breakup events (in yellow) produce higher breakup peak stages at the corresponding freeze-up level, consistent with the logic around ice cover strength and degradation. However, 2020-2021 stands out as a very high data point (significant peak breakup stage). When extending the data back to the 1996 freeze-up, as shown in Figure 4.15, the trend remains the same. However, the data point from 1996-1997 suggests that a worse-case scenario occurred in the spring of 2021, for hydrometeorological or thermomechanical reasons that need clarification.

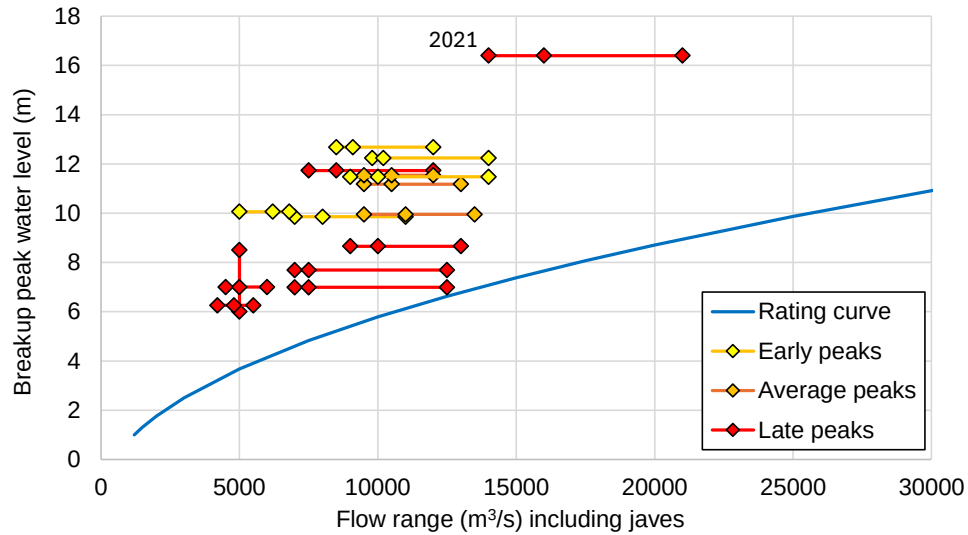


Figure 4.13. Possible range of flow associated with each ice jam (or ice cover) melt and mobilization events from 2011 to 2025 at station 10GC001. Color range is linked to cumulated degree-days of thaw (early = very few degree-days). The open water rating curve is presented for reference.

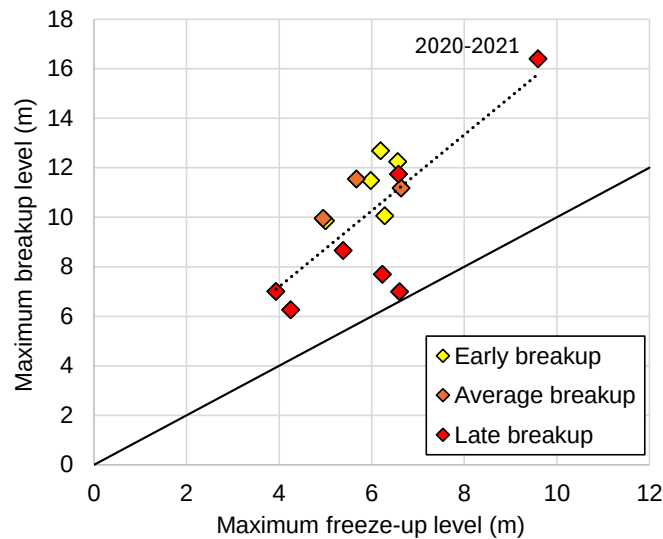


Figure 4.14. Maximum spring breakup level (mainly in the presence of ice jams) expressed as a function of maximum freeze-up level during the preceding fall at WSC station 10GC001 between 2011 and 2025. Colors represent the estimated ice degradation state based on cumulative degree-days of thaw.

Looking back at the project objective, is it possible that the concept of an absolute $Q_{\max \text{ ice}}$ applies to the Mackenzie River at Fort Simpson? Figures 4.13 and 4.14 advise that this is unlikely because $Q_{\max \text{ ice}}$ seems to vary (at least) as a function of the freeze-up stage (or intensity, or discharge), and this latter factor may not have its own absolute upper boundary. Also, the presence of isolated islands in the Mackenzie River could play a significant role in some years, providing a lateral outflow alternative and thereby removing some downstream hydrodynamic constraints near the toe of an ice jam. However, the two reported positions of the 2021 ice jam (Sentinel 1 on May 13, Figure 4.16; and NHC, 2025; ice jam toe 21 km downstream of Fort Simpson) do not seem to involve a nearby island.

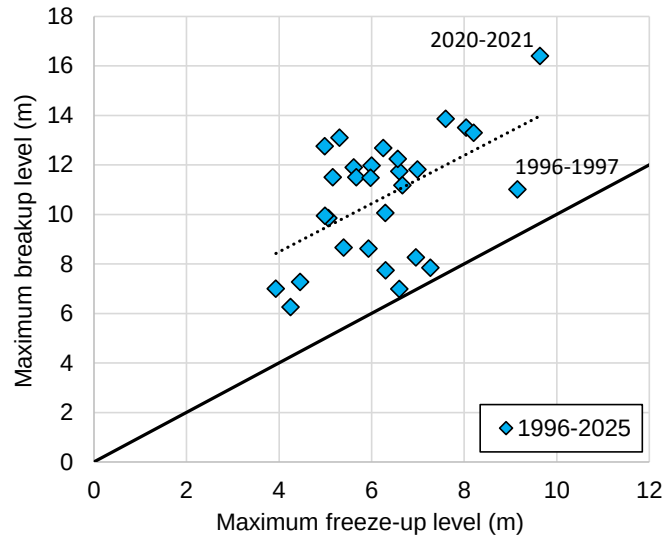


Figure 4.15. Maximum spring breakup level (mainly in the presence of ice jams) expressed as a function of maximum freeze-up level during the preceding fall at WSC station 10GC001 between 1996 and 2025.

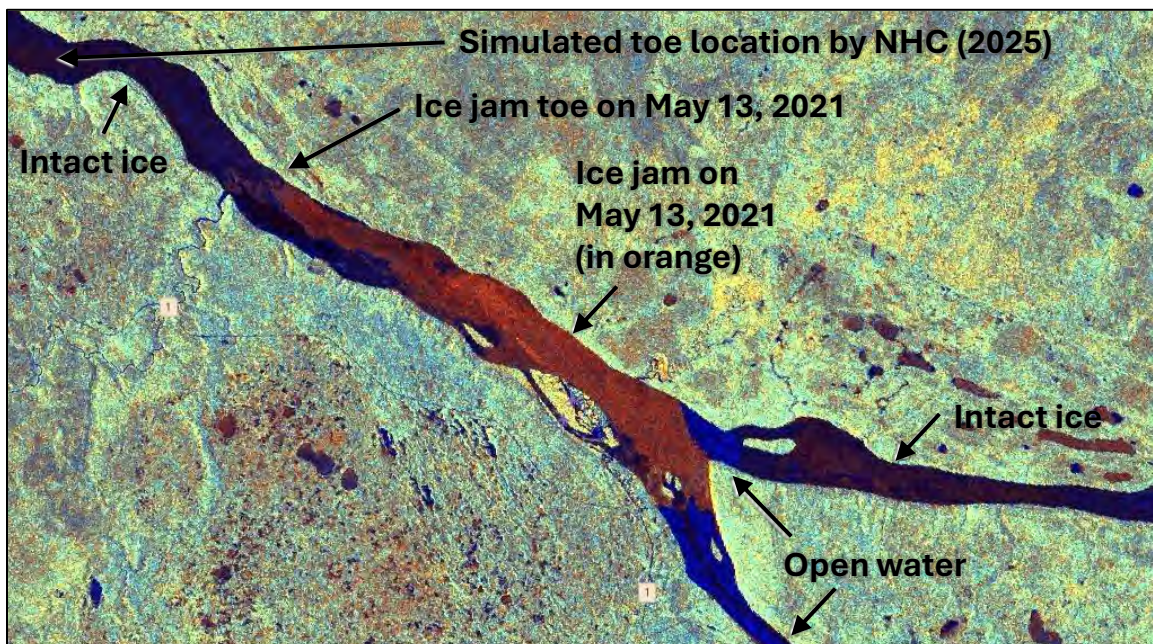


Figure 4.16. Sentinel 1 image (Copernicus, 2026) of the Mackenzie River at Fort Simpson on May 13, 2021, showing an ice jam at the community. The peak stage occurred on May 14.

One could also argue that erosion under the toe of an ice jam could contribute to releasing the pressure on the ice accumulation and therefore promote its stability despite the increasing discharge, which would challenge the concept of a fixed $Q_{\max \text{ ice}}$. The applicability of this theoretical sediment transport process to the 2021 ice jam is difficult to demonstrate. Finally, an interesting observation is presented in Figure 4.17: a significant multi-kilometre freeze-up consolidation in the Mackenzie River was present below the bend located 19 km downstream of Fort Simpson. This supports the hypothesis that Q_{ice} could be significantly impacted by an intense freeze-up. The post-freeze-up backscatter (i.e., surface roughness) was not as intense in any other year of the Sentinel 1 record.

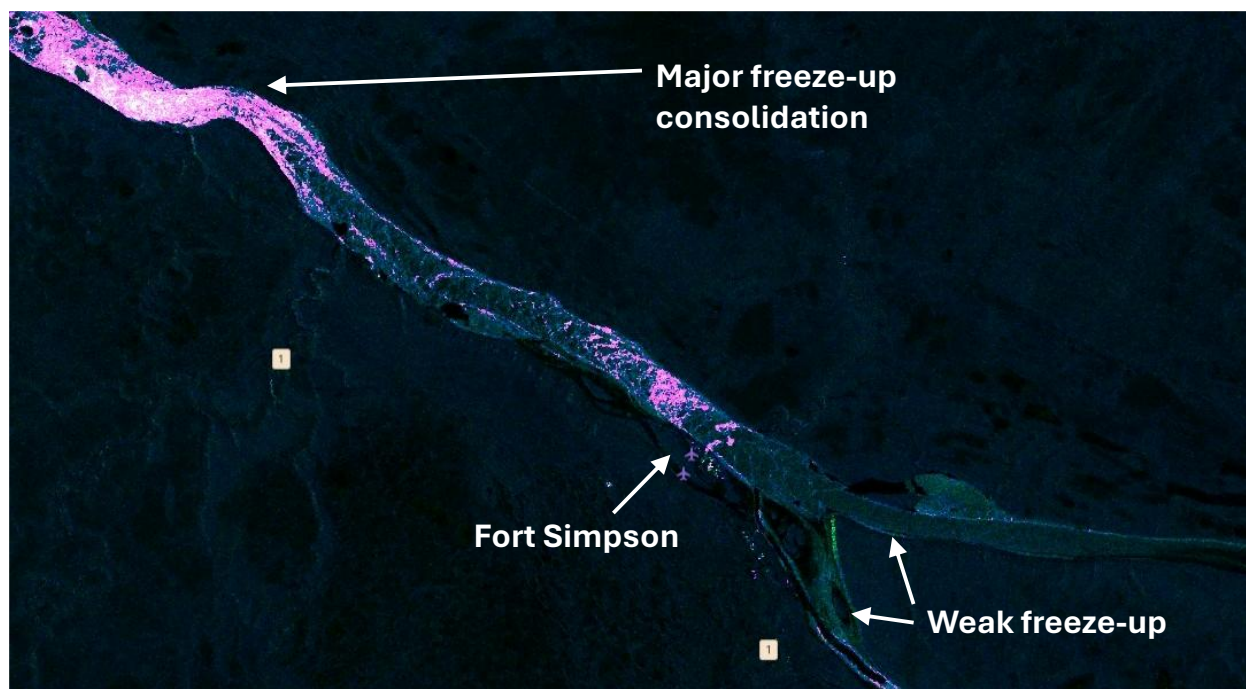


Figure 4.17. Sentinel 1 image (Copernicus, 2026) of the Mackenzie River on December 8, 2020, showing a significant freeze-up roughness (backscatter) starting 19 km below Fort Simpson and extending over 40 km downstream.

4.4. Chaudière River – Québec

The Chaudière River (Kikonteku or « river if the fields » in abénaqui) flows northward from the Appalachian Mountains near the US-Canada border to the St. Laurent River at the City of Québec in the province of Québec. It drains a watershed of 6,680 km². The hydrological regime of its headwaters is influenced by Mégantic Lake (27 km²). However, its middle reach is fed by several small, steep rivers, and this, combined with a high agricultural land-use ratio, contributes to its peaky hydrological behaviour throughout the year. The river ice dynamics of the Chaudière River have been described by Pelchat (2022), among others, a work now continued by Zidan et al., (e.g., 2025). Research on the winter regime of this river is justified by its frequent floods, including major ice jam events, as well as by its high exposure and vulnerability (several municipalities are established in the Chaudière River valley, including on floodplains). Table 4.4 presents hydrological and climatological data of the river near the Town of Beauceville.

Table 4.4. Hydrological and climatological information about the Chaudière River at Beauceville and St. Georges-de-Beauce in Québec.

Parameter	Approx. value
Average annual flow	75 m ³ /s
Average annual peak flow (Q ₂)	940 m ³ /s
Average annual minimal flow	5 m ³ /s
Average freeze-up date	15 Nov. – 15 Dec.
Average winter intensity	920°C-days
Average breakup date	25 Feb. – 10 Apr.
Date range of annual peak flow	Spring to early winter
Average annual air temperature	5.4°C

generate ice jams that remain frozen in place until spring. The thickness of these jams can be well above 1 m, which contrasts with the average ice cover thickness of 0.5 m (e.g., Pelchat, 2022; Figures 15-17), and this has obvious consequences on breakup resisting forces. Spring breakup events are not exclusively but generally driven by rain events in March or early-April. The reach downstream of Beauceville is usually the last to clear its ice cover and ice jams (Pelchat, 2022). Turcotte et al. (2020) indicate that the probability of ice jamming on the Chaudière River is higher in early-December (when the ice cover is still weak) and in late-March (when the driving force may increase sharply while the ice cover is at its thickest).

The discharge of the Chaudière River at Beauceville during spring breakup events was calculated by adding the estimated or measured discharge below the Sartigan Dam (Government of Quebec station 023429) with the measured discharge (generally based on the open water rating curve) of the Famine River (Government of Quebec station 023422). Water levels have been measured at Beauceville by the watershed organization (COBARIC) since 2015; therefore, only 10 spring breakup events were analyzed (gap in 2022). The Government of Quebec generally uses a backwater approach to produce winter discharge estimates, and this technique was considered reliable for the context of the current study, so very few flow corrections were applied to the publicly available 24-h or 15-min discharge time series. Figure 4.19 presents the peak breakup stage with corresponding estimated flow at Beauceville. The flow at which ice jams (or the ice cover) was mobilized is presented using black dots. It is of interest to note that in 8 springs (2022 is uncertain), the ice cover or ice jam at or below Beauceville released at a discharge that was lower than what had caused the maximum ice-induced stage. Looking at stage time series, it appears that the local ice jam or ice cover melted in place during 7 of 10 springs (the water level recedes gradually, whereas the incoming discharge follows a different trend). In turn, only 3 sudden ice cover or ice jam release likely occurred. This either means that the flow was not high enough to cause a dynamic release of the most significant ice jams (a runoff limitation), or that a $Q_{\max \text{ ice}}$ simply does not exist for that reach.

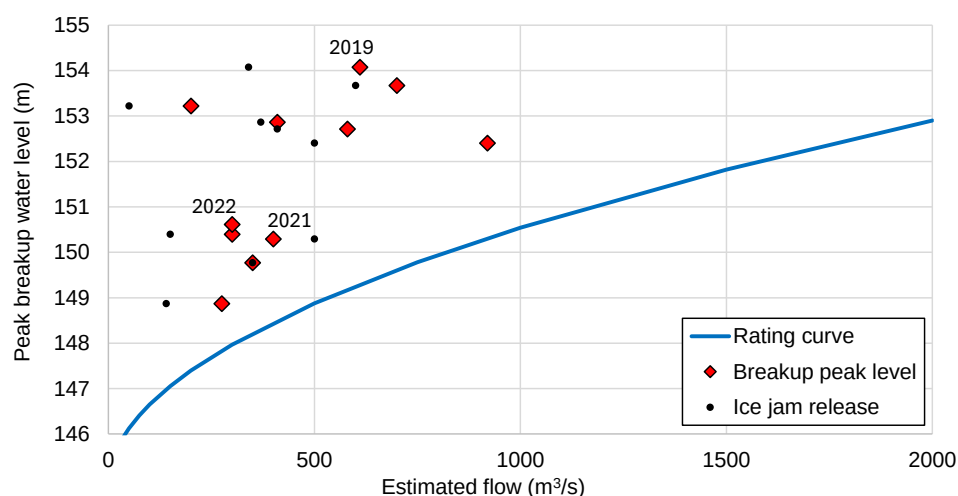


Figure 4.19. Open water rating curve (1969-2025 range) as well as peak and ice-jam release stage-discharge data from 2015 to 2025 on the Chaudière River at Beauceville, Québec.

The last phase of breakup seems to be dominated by thermal processes at Beauceville in most springs, including in the presence of major ice jams (e.g., 2019) that would have formed in a prior dynamic breakup phase. In this hybrid context, the impact of cumulative degree-days of thaw (CDDT)

on maximum ice-induced stage was investigated. This ice cover deterioration (resistance) indicator is occasionally used to anticipate the maximum intensity of spring breakup. Figure 4.20 presents the range of possible flows during each peak breakup event, complementing what is shown in Figure 4.19. In the case of the Chaudière River near Beauceville, an early breakup peak is associated with less than 15 CDDT whereas a late peak occurs at more than 30 CDDT (thresholds established so that breakup events are spread over categories). The results based on these thresholds send a misleading message: for instance, a low level of degradation can cause a small (2018 and 2023) or a significant (2016) backwater. In turn, the ice jams that resisted the highest flow occurred at a presumably higher state of ice degradation (2015 and 2019). The results from Figure 4.20 are not the most helpful to identify a possible $Q_{\max \text{ ice}}$ at Beauceville.

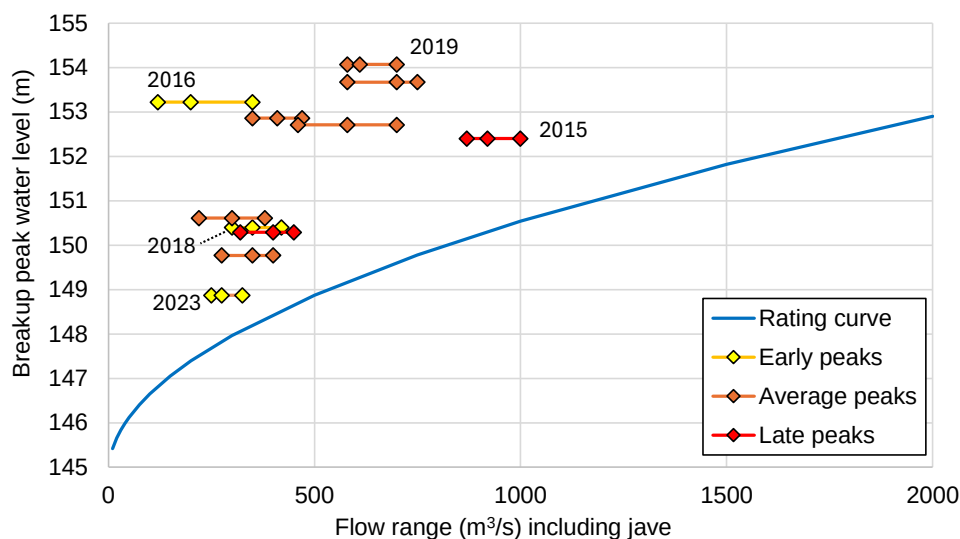


Figure 4.20. Possible range of flow associated with each ice jam (or ice cover) melt and mobilization events from 2015 to 2025 at Beauceville. Color range is linked to cumulated degree-days of thaw (early < 15 CDDT). The open water rating curve is presented for reference.

Building on a publication by Turcotte et al. (2025) on an investigation of the energy budget during the breakup period of three rivers in Canada, including the Chaudière River, rather than treating CDDT as an ice cover degradation indicator, a corrected or effective CDDT (ECCDT) was adopted. This gradually varying daily air temperature correction is meant to consider the effect of solar radiation on ice degradation beyond the impact of melting air temperatures, while also considering other factors such as cloud coverage and the presence of snow on the ice cover. The data in Figure 4.21 shows a slight category (colour) reshuffle compared with the graph in Figure 4.20. However, the results remain equally puzzling: Based on readjusted category thresholds (30°C-days or less for early-breakup events, 60°C-days or more for late events), data from 2024 is associated with a small backwater (a runoff-limited event) whereas a (assumed) deteriorated ice cover in 2015, 2019, and 2020 seems to have resisted the highest discharges. Those three outstanding events were further investigated:

- A significant runoff event in December 2014 could have either caused the formation of a mid-winter jam or a significant frazil consolidation near Beauceville. The winter of 2014-2015 was also much colder than usual, which could have generated a higher breakup resistance.
- Pelchat (2022) documented the presence of mid-winter ice jams downstream of Beauceville in 2019 (cold winter, major flood) and 2020 (average cold winter, minor flood).

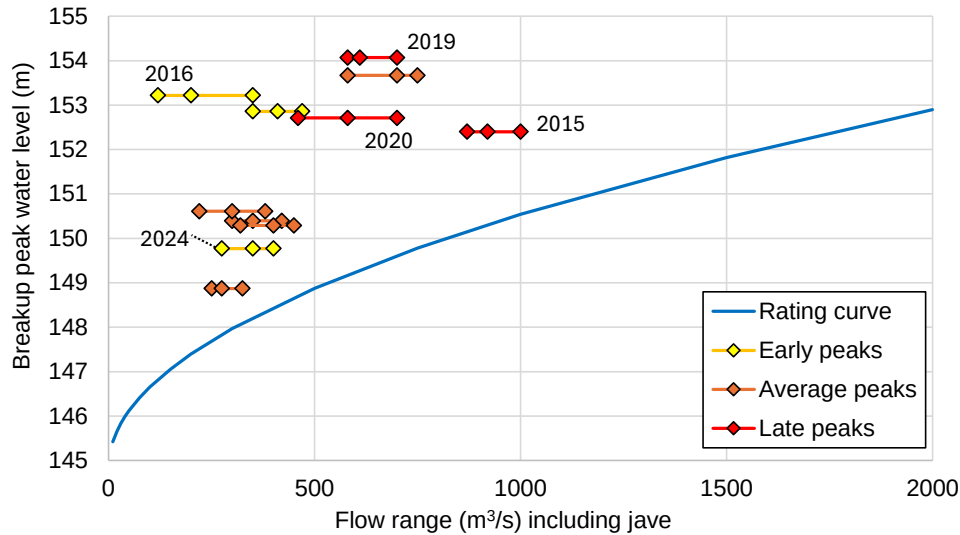


Figure 4.21. Same data as presented in Figure 4.20, but with a color range linked to effective or corrected cumulated degree-days of thaw (early = very few degree-days).

It is also of interest to note that the 2016 ice jam peak stage occurred at the end of February at 0°C-days (rain-driven). The ice jam resisted a moderate flow (it melted in place a month later) and could have likely generated higher water levels if the flow had continued to rise. That winter was surprisingly much warmer than usual.

In summary, it seems that 1. mid-winter ice jams, 2. significant post-mid-winter ice cover (frazil) consolidations, and 3. colder-than-normal winters have the potential to increase the resistance to spring breakup on the Chaudière River near Beauceville. The occurrence of a significant rain-on-snow event in late-winter or early spring, combined with one or many winter scenarios listed above, promotes the formation of larger ice jams, whereas dry spring conditions may prevent such an undesirable event. The great diversity of winter hydrological scenarios within the short, study period (only 10 years) makes it challenging to confirm or refute the possible concept of a $Q_{\max \text{ ice}}$ for Beauceville. This gets more complicated in a context where ice jam flood preventive measures are occasionally applied on the Chaudière River: municipalities often request the deployment of the Amphibex to break the ice cover before spring breakup, and the effectiveness of these interventions is uncertain.

Pelchat (2022) proposes that a $Q_{\max \text{ ice}}$ of 1200 m³/s may apply to the reach immediately downstream of Beauceville. This conservative value, as this author notes, also considers the state of ice degradation, calculated in degree-days of thaw above -5 (CDDT-5), a slightly different indicator than the ECDDT described by Turcotte et al. (2025). Pelchat (pers.com. 2026) explains that an absolute $Q_{\max \text{ ice}}$ may exist at some locations along the Chaudière River, but not downstream of Beauceville, where $Q_{\max \text{ ice}}$ would be a function of the ice degradation, mainly because of the low channel gradient combined with the presence of a wide floodplain.

5. Discussion

Table 5.1 summarizes the results presented in Section 4. Overall, the identification of a morphology or hydrological context in which the concept of an absolute, reach-specific maximum breakup discharge $Q_{\max \text{ ice}}$ could not be confirmed at any site. For each studied river reach, the breakup discharge during major ice jam events appears to be influenced or controlled by a downstream resistance factor, the nature of which varies from site to site.

Table 5.1. Key information about each studied river reach

River	Klondike	Hay	Mackenzie	Chaudière
Reach	Near WSC station 09EA003	Town of Hay River and delta (WSC station 07OB001)	Fort Simpson (WSC station 10GC001)	Town of Beauceville
Channel gradient	0.1% - 0.4%, some slope breaks	0.01% - 0.02% slope break in delta	0.01% - 0.02% Liard River is steeper	0.01% - 0.02% some slope breaks
Aerial pattern	Linear (low sinuosity), single channel, laterally confined	Linear, single channel in town, secondary channels in delta with some sinuosity	Confluence, single channel with occasional islands, low sinuosity	Single channel with occasional islands, one short rapid, some sinuosity
Floodplain	Limited	Mainly present in the delta	Limited	Significant
Breakup flow range	50 to 450 m ³ /s	30 to 1,800 m ³ /s	4,000 to 20,000 m ³ /s	100 to 1,200 m ³ /s
Annual peak flow (Q_2)	425 m ³ /s	610 m ³ /s	17,800 m ³ /s	940 m ³ /s
$Q_{\max \text{ ice}}$ dependency	Freeze-up jamming intensity	Great Slave Lake level during breakup	Freeze-up jamming intensity	Freeze-up or mid-winter jamming intensity

It is interesting to note that the upper limit of the breakup flow range is slightly above Q_2 for three of the four studied river reaches. This result alone represents an interesting observation that warrants additional exploration. In turn, it seems that Nature further tested the breakup limits of the Hay River in 2022 by producing an extreme runoff event during the spring breakup period, which coincided with an unusually high water level in Great Slave Lake (variables are independent of large climate patterns). The flow at the Town of Hay River during that event was at least twice the value of Q_2 (a channel-forming event from a fluvial morphology perspective) and could have momentarily reached three times Q_2 when taking incoming ice jam-release waves (javes) into account.

Another point of interest is that the 2023 ice jam on the Klondike River near Dawson and the 2021 ice jam on the Mackenzie River at Fort Simpson released dynamically as the flow increased (and the ice degraded). This means that the high resisting force threshold was exceeded by the increasing driving force. In turn, the 2022 ice jam on the Hay River at the Town of Hay River and the 2019 ice jam on the Chaudière River at Beauceville melted in place under rising and falling flows, respectively. For the

two latter rivers, this means that if a flow mobilization threshold for the ice jam exists from a hydromechanical perspective, it was not reached.

It is also interesting to explore climate patterns and weather conditions that led to the most severe ice jams presented in this work. In the case of the Klondike River and Chaudière River, relatively small watercourses, a sharp rise in air temperatures in the presence of a thick snowpack or a significant rain-on-snow event could produce flows much higher than Q_2 before ice clearance. This means that single-event, sub-regional weather patterns can lead to extreme ice jam floods in such watersheds. In turn, the ice jam floods that took place in the Mackenzie River and the Hay River depend, to a certain extent, on the water level in Great Slave Lake (either as a downstream hydraulic control or an upstream hydrological effect), and this is influenced by large-scale, multi-year patterns.

Climate and watershed-scale (e.g., aspect or orientation) influence on Q_{ice} or the concept of $Q_{max\ ice}$:

- In regions where river ice breakup is often triggered by rain-on-snow events, there is less of a limit on the flows that can materialize in the presence of ice, and therefore the $Q_{max\ ice}$, if it exists, would be easier to document. For instance, many river ice breakup events in the Chaudière River, a watercourse that may be representative of rivers in Southern Ontario, Southern Québec, and New Brunswick, occur with essentially no degree-days of thaw and therefore minimal ice degradation.
- In regions where river ice breakup is generally driven by snowmelt under dry conditions, the flow during breakup may be limited by the rate at which the snow turns into water (or how warm air temperatures can be in a snow-covered watershed where ice is still present in rivers). Since the ice cover is also affected by warm air temperatures, it would be more difficult to document a possible $Q_{max\ ice}$. This applies to large rivers of Canada's Northwest.
- In large, northward-oriented watersheds, even those with a dry, snowmelt-dominated freshet, estimating a $Q_{max\ ice}$ would be possible because, in some, if not most years, the ice cover may not be significantly degraded downstream, whereas snowmelt runoff begins in the (often warmed) south.

The results presented in this study have limited generalizability for a few reasons:

- Few river reaches were included (the most documented sites with a recent major ice jam event were prioritized), which limited the diversity of studied cold region river morphologies.
- The analyzed data series for each site was short (limited by the availability of instantaneous stage data and satellite imagery).
- There were several gaps in hydrometric station records (often a consequence of dynamic breakup events), which translate into significant uncertainties. The Fort Simpson station (Water Survey of Canada 10GC001) and neighbouring stations were affected by several gaps during breakup, and the Athabasca River was excluded from this study due to consistent gaps in hydrometric records at breakup.

Nonetheless, the results presented in this work, especially in Figures 4.4, 4.9-4.10, and 4.14-4.15, open an interesting investigation path into reach-specific (morphology-based) variable $Q_{ma\ ice}$, which would be a function of the probability of ice-related parameters controlling the resistance (force) to breakup. It would merge the material presented by Pelchat (2022) and Turcotte et al., (2024) on breakup thresholds for different river reaches, and the site-specific flood mapping approach developed by Healy and Kovachis (2023) about the probability of different breakup scenarios. For

example, Figure 22 in Pelchat (2022) is replicated below with modifications (Figure 5.1). This graph presents the streamwise variability in upper and lower discharge thresholds at which the ice cover or ice jam is mobilized along the Chaudière River (the upper boundary, to some extent, being equivalent to $Q_{\max \text{ ice}}$). A similar concept had been proposed by Burrell et al. (2022; see Figure 5), with the original idea depicted by Beltaos (2008, see Figure 6.13). Here, it is associated with different channel characteristics (e.g., gradient, planform).

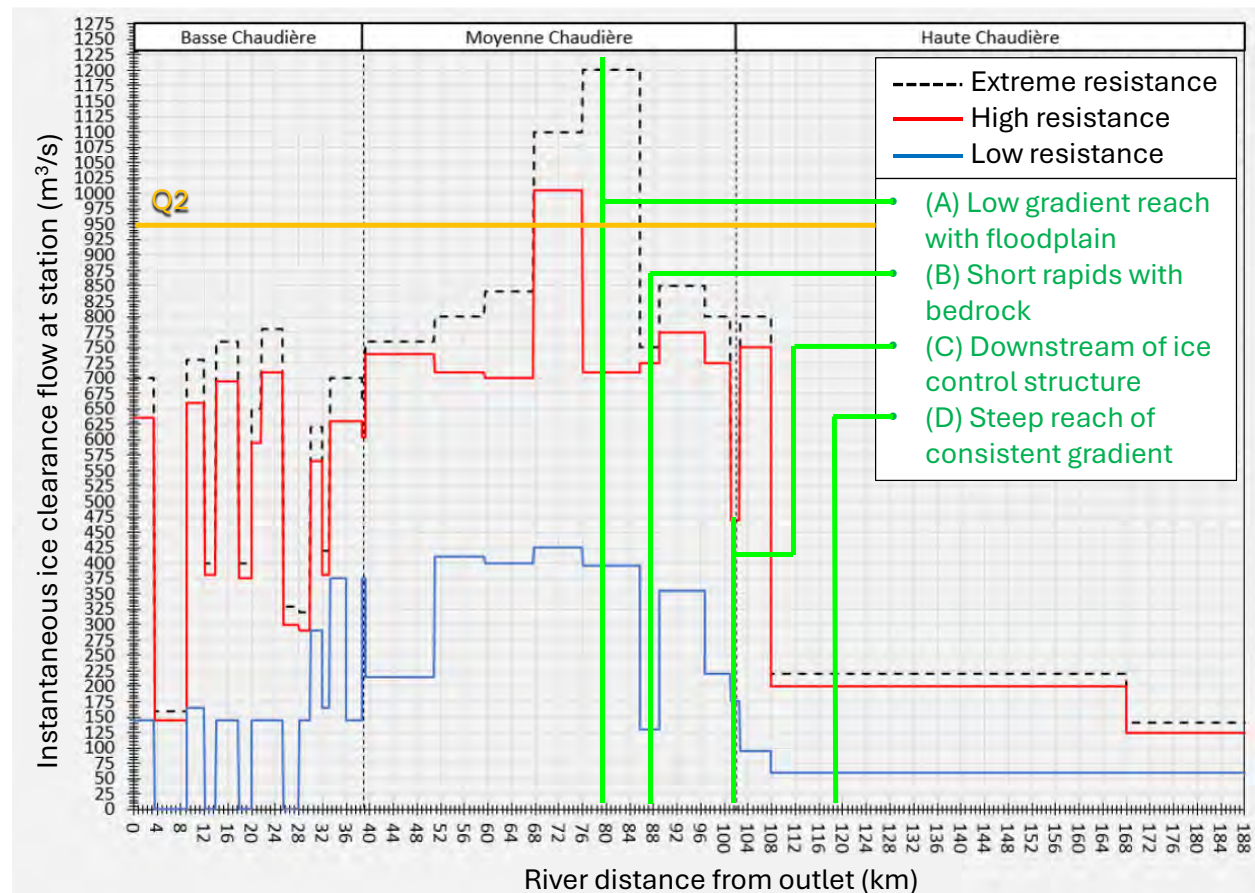


Figure 5.1. Reproduction of Figure 22 in Pelchat (2022). Lower, upper, and extreme breakup flow boundaries expressed as a function of river kilometers upstream the Chaudière River.

In Figure 5.1, possible ranges of mid-winter or spring breakup flow thresholds are presented for different river segments:

- Line A presents the breakup flow range for the low-gradient reach of the river, which includes a wide floodplain. Mid-winter jams often form in the reach. In this case, the upper boundary ($Q_{\max \text{ ice}} = 1200 \text{ m}^3/\text{s}$) is greater than Q_2 ($940 \text{ m}^3/\text{s}$).
- Line B presents an equivalent flow range for a short rapid where the ice is mobilized at a relatively low flow. This may be explained by three factors: 1. More friction heat causing an early ice melt, 2. More fragile ice mobilized at a lower flow, 3. Short ice jams causing a rapid increase in the driving force.
- Line C presents a breakup flow range that is influenced by the absence of ice supply from upstream (including frazil that would otherwise transit down the river during the first weeks of

winter) because of the presence of a dam. The thresholds at Line C are therefore lower than what naturally occurs without the presence of the dam.

- Line D presents the possible range of mid-winter or spring breakup flow thresholds for the steep, upstream reach of the Chaudière River. In this case, the identified flow breakup range ($60 \text{ m}^3/\text{s}$ to $225 \text{ m}^3/\text{s}$) is much lower than Q_2 , and the resulting ice run is intercepted by the dam.

It is worth returning the reader to Figure 3.1 of this report. Zone 1 (relatively low breakup resistance) may apply to the location defined by Line D in Figure 5.1, whereas Zone 3 may correspond to Line A. The concept of a reach-specific breakup flow range for annually variable ice cover resistances is important because there is sufficient information in the river ice literature to build equivalent data sets and to link the identified thresholds to specific morphologies which could, in turn, inform future studies (e.g., flood mapping).

The following step is to link these “fixed” thresholds to actual probabilities. Figure 5.2 emphasizes lines A and D in the graph presented in Figure 5.1. The next question would be how to define these probabilities? The answer is that it would require several years of historical measurements and observations, just like what Healy and Kovachis (2023, and NHC, 2025) did by calculating the probability of breakup scenarios by considering a range of flows associated with thermal and mechanical breakup events for a specific site.

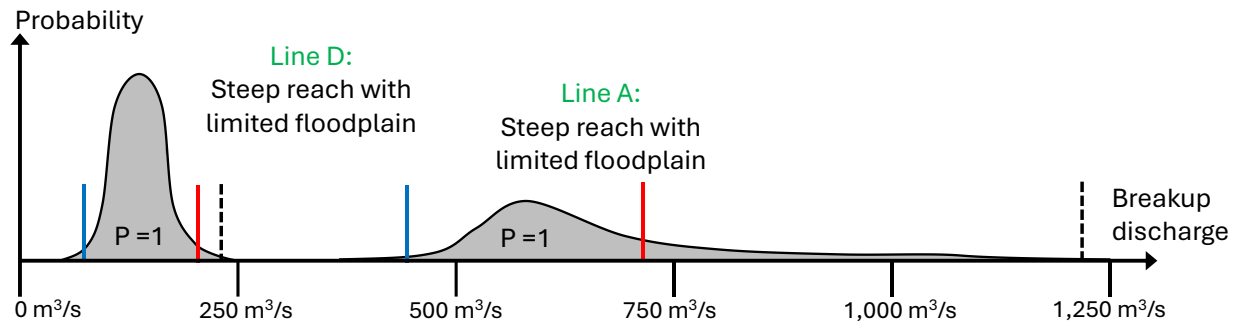


Figure 5.2. Conceptual graph presenting the probability distribution of ice cover or ice jam breakup flows for two different reaches of the Chaudière River, based on Pelchat (2022).

Figure 5.3 extends this logic a bit further, using the example of Figure 4.4 above for the Klondike River and associating the two main variables that seem to affect resistance to breakup with ballpark probability distributions. It is important to note that the probability of both parameters (freeze-up intensity and state of degradation of the ice cover at the time of mobilization) is fairly independent, but this may not always be the case. The main parameters controlling the resistance of an ice cover or ice jam to breakup logically vary from site to site, and it is for the scientist, engineer, or researcher to identify these controls.

The conceptual figures 5.1 to 5.3 can help guide the development of flood maps in long river reaches where breakup thresholds may vary. Moreover, by taking the morphology of different channel segments into account and by considering possible breakup sequence scenarios, they can represent a framework for the development of river ice breakup forecast models and for the design of hydraulic structures with or without the use of a hydrodynamic model (that requires the costly acquisition of bathymetric data). If a hydrodynamic model is available or deemed affordable, defining reaches with distinct breakup dynamics can inform the probability distributions of different ice jam parameters

included in a Monte Carlo approach. In other words, even if an absolute, reach-specific $Q_{\max \text{ ice}}$ may not exist in any cold region river context, a breakup Q_{ice} range (with associated probabilities) relative to Q_2 may be associated with specific morphologies or morphology sequences.

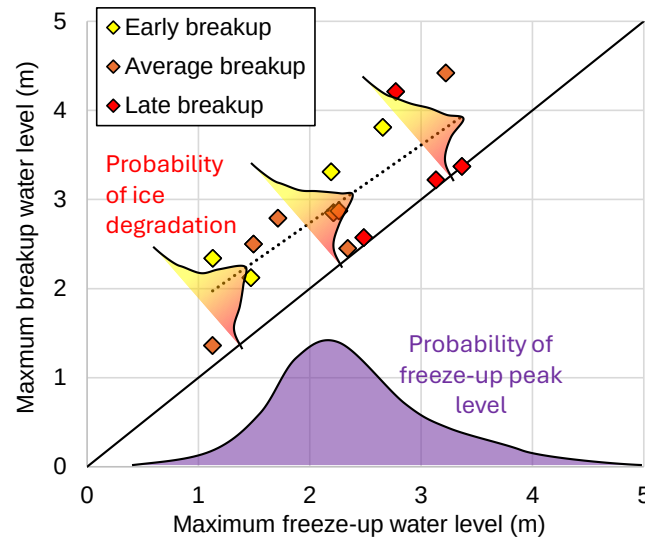


Figure 5.3. Approximate probabilities of two variables impact maximum breakup stage on the Klondike River at station 09EA003 operated by the Water Survey of Canada.

It is important to recall that simply adopting post-ice clearance flows (last B date for WSC historical record) to represent hydrological conditions causing peak ice-induced water levels along small and long river reaches may lead to inaccuracies varying from minor to hardly defensible from a technical point of view (Figures 5.4-5.5).

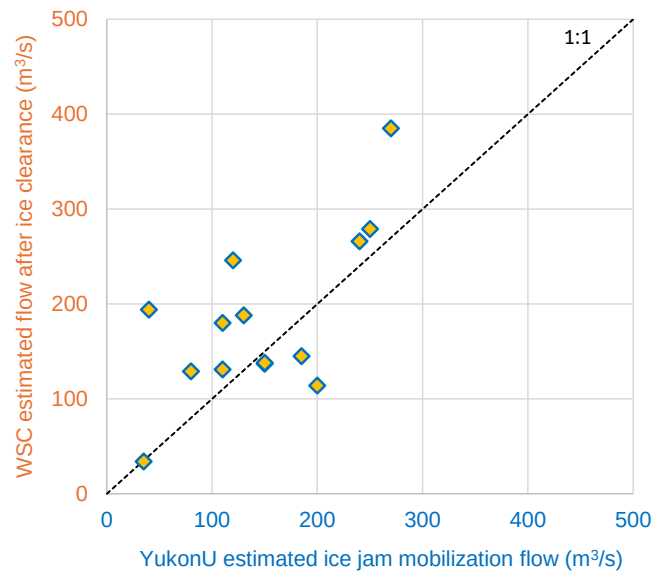


Figure 5.4. Post breakup flows (derived from a rating curve) expressed as a function of reassessed (see Section 3) flow during peak stage for the corresponding breakup event for the Klondike River (data from 2011 to 2024).

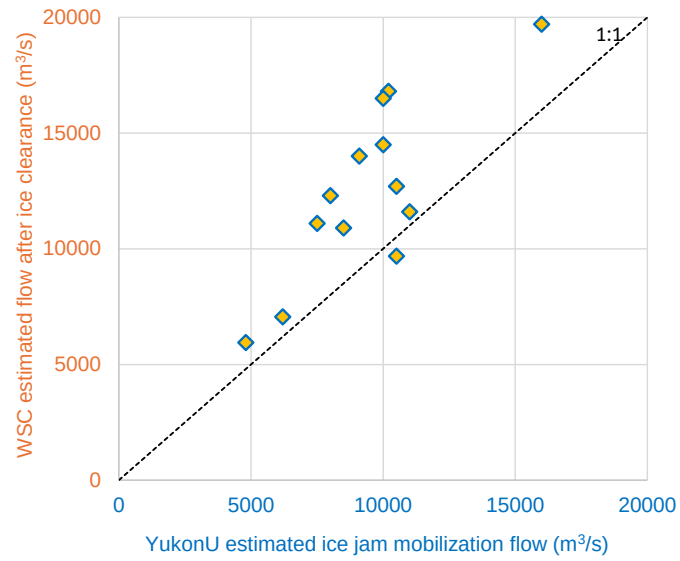


Figure 5.5. Post breakup flows (derived from a rating curve) expressed as a function of reassessed (see Section 3) flow during peak stage for the corresponding breakup event for the Mackenzie River (data from 2011 to 2024).

6. Proposed research framework

This report has opened several potential research avenues about river ice breakup in cold regions and its implications for scientific and engineering projects. One important objective of this work was to propose a research framework to explore the existence and applicability of $Q_{\max \text{ ice}}$ across distinct cold region channel morphologies in flood mapping, infrastructure design, and ice jam flood forecasting contexts. The results presented in Sections 4 and 5 (no absolute $Q_{\max \text{ ice}}$, but an obvious dependency on other parameters) force an adaptation of this objective.

- Practitioners are invited to re-assess streamflow during the peak breakup stage for critical projects:
 - The methodology described in Section 3 and presented in more detail in Turcotte and Rainville (2022) is probably worth using to verify or re-assess historical flow time series during the breakup period, more specifically during the instant of the peak breakup stage. This statement applies to most scientific and engineering projects.
 - It is rarely adequate to associate post-breakup flows to the maximum breakup (or ice jam) stage. This is demonstrated in Figures 5.4 and 5.5, and it has direct implications for engineering projects (less so for flood forecasting, because the open water rating curve does not apply until breakup is over).
 - It is important to include as many breakup events as possible in the analysis of each river site or segment, ideally including different scenarios (e.g., early and late breakup events, thermal and mechanical events). The CRID database (Government of Canada, 2020) could be used to expand the current project to additional ice-affected river sites in Canada, including those where ice jams do not represent a dominant flooding process.
- This report has mainly covered river segments (and associated morphologies, presented in Table 5.1) linked with high/extreme breakup thresholds, three of which included a community established on the floodplain. In this context, it would be valuable to identify cold region river morphologies associated with very distinct breakup resistance controls, and, therefore, with different ranges (or probability distributions) of river ice breakup flow relative to a reach-specific hydrological indicator (e.g., the average annual peak flow, or channel-forming flow Q_2). The following list is by no means exhaustive, but may be useful for an initial exploration:
 - Rapids with emerging rocks (or bedrock) and a limited floodplain (low breakup flow threshold)
 - Deep, fast-flowing channel sections in an incised valley (i.e., canyons, low threshold)
 - Low-gradient rivers with islands and a wide floodplain (high threshold)
 - Low-gradient rivers with vertical banks and no floodplain (moderate threshold)
 - Small, low-gradient meandering rivers (high threshold)
 - Anastomosed rivers with a low floodplain (high threshold)
 - Wide braided river segments in subarctic environments (no breakup threshold)
 - Riffle-pool channels with occasional floodplains (moderate threshold)

Include stage-flow data from these various hydro-morphological contexts into a graph similar to what is presented in Figure 2.6 (with dimensionless axes).

- Explore the concept of a multi-reach (or breakup sequence) analysis as presented in Figure 5.1, not necessarily involving fixed thresholds ($Q_{\max \text{ ice}}$), but potentially 5% and 95% flow threshold boundaries.

In addition, considering the results presented in this report that are beyond the objectives of the projects, relevant complementary river ice research could consist of:

- Develop simple probabilistic models to determine a possible range of breakup water levels based on the knowledge about the dominant ice cover or ice jam resisting controls (i.e., present non-conceptual figures similar to Figure 5.3). The results could be compared with a direct probability distribution of measured peak breakup water levels or with those of uncalibrated and calibrated Monte Carlo simulations.
- Investigate the role of javes on breakup sequences and breakup flows. Flow instabilities during river ice breakup events are always present, and it is possible that most ice jams release due to small or large javes rather than solely because of a gradually increasing driving force or a gradually decreasing resisting force. When this occurs, it theoretically (from a mobilization force perspective) means that the ice jam could have resisted a higher flow (driving force) in the absence of any jave.
- Investigate the impact of climate change on the probability distribution of site-specific breakup flow ranges across different rivers in Canada. It is often assumed that climate change will simply result in a shorter ice season and thinner or ephemeral ice covers in cold region rivers. It is, however, generally difficult to justify the impact of climate change on the risk of ice jam floods without a profound understanding of site-specific breakup controls and sequences.
- Investigate how river ice processes may influence a river's morphology. This means that a positive feedback loop between channel characteristics and breakup thresholds may exist (e.g., Turcotte et al., 2011, Turcotte et al., 2013, Nafziger et al., 2026). This could also mean that ice jams and their release represent channel-forming events (just like Q_2) or channel-destabilizing events (just like extreme open water runoff events).

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References

- Beltaos, S. (2008). River ice breakup. Water Resources Publications, Highland Ranch, CO.
- Burrell, B.C., Beltaos, S., Turcotte, B. (2022). Effects of climate change on river-ice processes and ice jams, International Journal of River Basin Management, DOI:10.1080/15715124.2021.2007936
- Government of Canada (2020). Canadian River Ice Database. <https://open.canada.ca/data/en/dataset/c5b58ccd-0011-4a80-8f24-034c86cbc14d>
- Healy, D., Kovachis, N. (2023). Case Studies of Ice Jam Flood Frequency Estimation. 22nd CGU-HS CRIPE Workshop on the Hydraulics of Ice Cover Rivers. Camore, AB. July 9-12.
- Kovachis, N. (2011). Patterns of River Breakup Timing and Sequencing, Hay River, NWT. Master's thesis. University of Alberta, Edmonton, AB. 459 p.
- Ladouceur, J.-R. (2022). Détermination des cotes de crues en présence de glace sur la rivière Chaudière. Mémoire de Maîtrise, Université Laval.
- Lindenschmidt, K.-E. (2017). RIVICE—A Non-Proprietary, Open-Source, One-Dimensional River-Ice Model. Water 9, 314.
- Nafziger, J., Turcotte, B., Berti, Z. (2026). The Interdependence of River Ice and Geomorphology: An Invitation to Conversation. Oral presentation. CGU-IAH-CNC Conference, Halifax, NS, May 24-27.
- Natural Resources Canada (2023). Federal Hydrologic and Hydraulic Procedures for Flood Hazard Delineation, Version 2.0. 169 pages. <https://doi.org/10.4095/332156>
- NHC (Northwest Hydraulics Consultants Ltd.) 2025. Fort Simpson Flood Hazard Mapping Study. Prepared for the Government of Northwest Territories. 147 p.
- Turcotte, B., Pelchat, G., Ghobrial, T. (2025). Correction of the degree-day approach to take solar radiation into account in breakup forecast models. 23rd CGU-HS CRIPE Workshop on the Hydraulics of Ice Covered Rivers, St. John's, NFL, June 9-12.
- Turcotte, B., Saal, S., Dubnick, A., Horton, B., Zammit, A. (2024). Developing new river ice breakup forecasting tools in the Yukon – Klondike River near Dawson. Presented to the Water Resources Branch, Department of Environment, Government of Yukon. YukonU Research Centre, 52 p.
- Turcotte, B., Rainville, F. (2022). A new winter discharge estimation procedure: Yukon proof of concept, in 26th IAHR International Symposium on Ice, Montréal, QC.
- Turcotte, B. (2022). Improving winter discharge estimates, Phase II – proposed new procedure and proof of concept. Presented to the National Hydrological Services, Meteorological Services Canada. Prepared by the YukonU Research Centre, Yukon University, 57 p.
- Turcotte, B., Saal, S., Horton, K. (2021). Preliminary assessment of flood exposure for future development areas in Dawson and Carmacks. Presented to the Land Development Branch, Department of Community Services, Government of Yukon. YukonU Research Centre, Yukon University, 42 p.
- Turcotte, B., Morse, B. (2015). River ice breakup forecast and annual risk distribution in a climate change perspective. 18th CGU-HS CRIPE Workshop on the Hydraulics of Ice Covered Rivers. Quebec City, August 18-20.
- Turcotte, B., Morse, B. (2013). A global river ice classification model. Journal of Hydrology 507: 134-148.
- Turcotte, B., Morse, B., Bergeron, N.E., Roy, A.G. (2011). Sediment transport in ice-affected rivers. Journal of Hydrology 409 :561-577.
- Zidan, E., Ghobrial, T., She, Y. (2025). Forecasting river-ice breakup timing using hydrometeorological and morphological parameters: a case study of the Chaudière River. 23rd CGU-HS CRIPE Workshop on the Hydraulics of Ice Covered Rivers. St. John's, NFL. June 9-12.