Notes on

I C E J A M S

by

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INTRODUCTION

Without doubt the most dramatic event on a northern river is the formation of a large ice jam. This can cause water levels that far exceed even the largest summer, or open water, flood levels, with obvious consequences for riverside communities and engineering structures. Figure 1(a) compares breakup and summer flood levels for Fort Vermilion on the Peace River in Alberta. The location is shown in Figures 1 and 2 of Appendix II. The dominance of the breakup water levels is obvious. The view from the front door of one of the riverside homes in the town during the fourth highest flood is shown in Figure 1(b). Bridge superstructures must obviously be placed well above such levels to avoid the problems shown in Figures 2 and 3, and development located to avoid the problems shown in Figures 4 and 5.

The sudden failure of ice jams can cause high velocity flow and the movement down river of large ice floes at high water levels. It is noteworthy that each pier of the bridge recently constructed at Fort Vermilion was designed to resist the full ice load of 7 MN applied at the highest breakup stage shown in Figure 1(a). Ice jams can also cause unusual scour both of the bed and banks, the latter more by the flow of water in unexpected locations rather than the physical abrasion of the ice.

Ice jams are therefore an extremely important feature of river engineering in cold regions*. Yet, in comparison with summer floods, their characteristics are poorly known.

Ice jams can be very local and very brief, yet very damaging. In unpopulated regions they are also unrecorded. These features make it desirable that the mechanics of ice jam formation and behaviour be understood because statistical records of breakup water levels are few and, more importantly, unlike summer flood records, those few cannot be transposed to other locations along even the same river.

ICE JAM TYPES AND CHARACTERISTICS

Ice jams can be broadly classified on the basis of the season in which they form - freeze-up, winter and breakup - and of their type - floating and grounded.

Freeze-up Jams

These form when the stream becomes gorged with frazil ice, as shown in Figure 6, or when the down-river passage of pancake ice becomes obstructed and a jam forms.

Winter Jams

These form when a mid-winter thaw causes breakup. By definition such a breakup does not extend over a long length of the stream. The supply of

* When defining the geographical limits of cold regions it is well to recall events such as the ice jam in 1899 on the Mississippi River at New Orleans! (Gerdel, 1969).

ice floes is therefore limited and the increase in discharge is of short duration. These two features generally limit the magnitude of the water level increases. The major significance of such jams is that they refreeze forming a formidable obstruction for the subsequent spring breakup. This is also a danger with freeze-up jams (for example, see Frankenstein and Assur, 1972).

Breakup Jams

Generally these are of most concern and form during the general spring ice run.

After initiation an ice jam can develop into a floating or grounded ice jam.

Floating Jam

This type of jam maintains a relatively unobstructed flow of water under its full length, except perhaps for a short section near the toe (downstream end) of the jam. It seems to be the most common type of jam and is sketched in Figure 7(a).

Grounded (or Dry) Jam

In this jam type the ice accumulation extends to the stream bed over a considerable portion of the length of the jam. The jam then behaves much like a rockfill dam, as shown in Figure 7(b), with the character of the flow being that of flow through porous media. High water levels can therefore be expected.

The discussion that follows deals with breakup jams. Such jams will obviously depend heavily on the time and manner of breakup. This is briefly reviewed first.

BREAKUP AND ITS PREDICTION

First, it is important to realise that there are some rivers in cold regions which rarely, if ever, experience a well-defined ice run. Such streams are generally braided and shallow with large expanses of ice frozen to the bed, such as the Delta River shown in Figure 8 (which is nowhere near a delta). Such streams are very common in N.W. North America. However for streams in which an ice run is a regular feature, the nature of breakup at a given location depends on:

(i) snow melt (magnitude and rate of rise of water level);
(ii) thickness and strength of the ice cover;
(iii) water level at freeze-up;
(iv) quantity of ice moving down from upstream and, last, but definitely not least;
(v) morphology of the river.

Breakup can progress upstream or downstream depending on the orientation of the river and its tributaries relative to the spring isotherms and the occurrence of snowmelt and/or spring rains. In many instances breakup occurs first along the central portions of a stream because of the breakup of a major tributary. However, no matter in which direction breakup progresses, it is a progression only in a very general sense; there are many local perturbations, these often taking the form of major ice jams.

Breakup is instigated by changes in one or both of two features: water level and ice sheet strength. The ice can become so weak that a low flow is sufficient to fragment and move the ice out. In this case the ice run will be minor. At the other extreme the water level and flow can increase sufficiently to float a strong ice sheet free of the bed and banks and to fragment the ice sheet. For a competent ice cover it would seem breakup can only occur in an intermittent fashion, with ice jams forming, however briefly, to build up water levels and release surges. Such a surge will move ahead of the fragmented ice to keep the breakup front moving. As will be discussed later, the celerity of such surges can be very high.

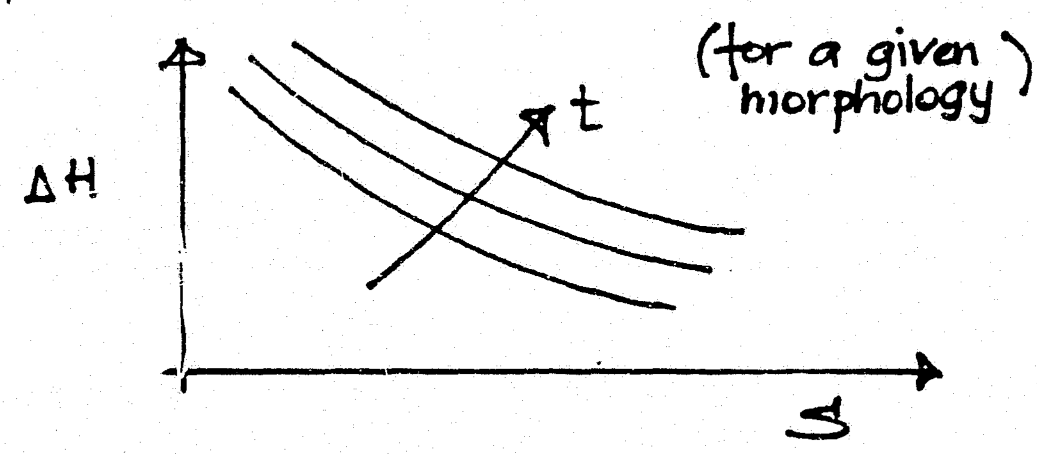
From the above discussion it would seem the three most pertinent parameters governing the moment and manner of breakup at a given location are:

(i) the difference in water level from that just after the formation of a stable ice cover during freeze-up, ΔH;
(ii) ice thickness, t_i;
(iii) the number of degree days of thaw, S, which provides a measure of the ice strength.

That item (i) is relevant is supported by the graphs shown in Figure 9, taken from Shuliakovskii (1963). Item (iii) is supported by the other graph shown.

If item (iii) is important there is little doubt that ice thickness should also be a parameter, although it may vary little from year to year at a given site. However, it should be remembered that the natural ice thickness can be modified by the formation of a freeze-up jam, winter jam or aufeis.

Presumably, for a given river morphology, the relation between breakup and these parameters will have the form:



Unfortunately no systematic evaluation of such a function has been reported in North America. About all that can be said at present is that breakup will not occur until about 30°C days have accumulated and the water level has increased somewhat beyond that at freeze-up.

The required increase in water level can be caused by snowmelt (or rain) or by an ice jam failure upstream. Either of these can occur on the mainstream or an upstream tributary.

To give some idea of the way breakup progresses Figure 10 shows a summary of the average breakup dates for the major streams in Alberta, Canada. There are several features of interest. As mentioned above, several streams breakup in their central reaches first, the breakup generally being triggered by breakup in a major tributary. This role of tributaries in causing breakup on the mainstream can be an important consideration. If the relative discharges of tributary and mainstream are changed (for example, by regulation or diversion) this will change the influence of the tributary on breakup in the mainstream, and, consequently, may change the frequency of ice jams at and near the confluence.

Also of interest in Figure 10 is the concentration of the isochrones at the Wabasca - Peace confluence near Ft. Vermilion. This is probably indicative of ice jams at the confluence and suggests that, unlike the Smoky River near Peace River town, breakup on the Wabasca is not strong enough to cause breakup on the Peace River.

In addition to being important in the spring, the risk of inducing breakup imposes important constraints on the allowable range of discharges from hydro-plants (Burgi et al., 1971; Pentland, 1973).

Some field observations of breakup have been reported (eg. MacKay, 1965; Newbury, 1967; Johnson and Kistner, 1967; Nuttall, 1970; Slaughter and Samide, 1971; Sampson, 1973; McFadden and Collins, 1977); and Michel and Abdelnour (1975) have done some preliminary studies in the laboratory using simulated ice, but in general much of engineering significance remains to be learned about the common event called breakup.

INITIATION OF ICE JAMS

The initiation of ice jams during breakup will probably be a function of the same variables as breakup. Hence ice jams can be expected with large ice thickness, heavy snow accumulation and a large and rapid increase in temperature above freezing. On the other hand, ice jams are less likely if there has been little snow or there is a gradual onset of spring. If an ice jam forms its severity will be a function of the rate of rise of water level (and the associated velocity), the amount of ice travelling with the breakup front, and the nature of the obstacle that initiates the jam.

Obviously with all these parameters fixed, the probability of a jam at a particular location will depend on the river morphology. This can at least be roughly analysed. Using a simple analysis Nuttall (1973) has shown that locations of large mean depth, relative of the average for the stream, cause an increase in concentration of floating ice and hence constitute an hydraulic obstruction to the passage of ice and increase the chance of jam initiation. Such locations will correspond to bends and narrows. At these locations, the plan form of the stream provides a further impediment to the passage of ice. The headwaters of reservoirs provide other examples and these are indeed a common location of ice jams. Sudden changes in slope from steep to flat also seem to be prime ice jam locations. This is presumably related to the deepening of the flow and the decrease in velocity.

Such high ice concentrations can also be caused by physical obstructions such as islands and bars. A very common physical obstruction is the ice sheet on the river, particularly if it is thick or more shore - or bottom - fast than normal (eg. because of freeze-up of winter jams, aufeis, or hanging dams). The mainstream ice cover is a frequent cause of ice jams at the mouth of tributaries.

The Athabasca River at Fort McMurray, Alberta, (see Figure 10), is an example of a location where both hydraulic (sudden decrease in slope and increase in depth and width) and physical obstructions (islands, bars, bend, wide ice sheet) exist. Not surprisingly, therefore, it is a location where ice jams form almost annually.

FLOATING ICE JAMS

Ice jam initiation has been briefly discussed. After initiation the future characteristics of a breakup jam depend on a series of hydraulic and structural constraints.

A general analytical framework for determining the major characteristics of floating ice jams was established over a decade ago by Pariset and Hausser (1961), Michel (1965) and Pariset, Hausser and Gagnon (1966). The analysis reflected that of an earlier investigation of the mechanics of log jams by Kennedy (1958). Some refinements to the analysis were added by Uzuner and Kennedy (1976) but, as pointed out by Beltaos (1978), the essentials remained unchanged. The latter investigator applied the analysis to two natural ice jams on the Smoky and Wapiti Rivers in Alberta with encouraging results. Another successful application is reported by Macdonald and Hopper (1972). Although further confirmation under field conditions is obviously desirable, the approach seems viable. It should therefore be possible to determine reasonable values for the maximum breakup water levels at a site caused by steady floating ice jams, using records of past breakup discharges. The latter are generally both transposable and available.

Hydraulic Constraints

If the surface velocity is low enough ice floes will simply accumulate against the solid ice cover or obstruction and the accumulation front will move upstream to leave behind an accumulation one layer thick.

However, if the velocity exceeds a critical value the ice floe will turn under when it contacts the obstruction, and will be entrained by the flow. Again depending on the flow velocity, it may then be deposited under the downstream ice cover or carried on downstream.

The deposition velocity has yet to be investigated in any detail for freeze-up or breakup conditions. If the latter occurs the ice front cannot move upstream until some event occurs downstream to lower the velocity near the front. If the former occurs, floes will accumulate under the downstream ice cover until the obstruction is such that the surface velocity at the front is reduced and the impinging ice floes are not entrained. The front will then begin to move upstream and the process of entrainment, deposition and surface accumulation repeated. This will result in a steady progression of the front if another condition is satisfied.

It is argued that there is a tendency for the front of such an accumulation to be entrained by the flow as it moves under the accumulation. This requires a local acceleration of the flow which in turn requires a lower piezometric head downstream. This causes a lowering of the water level just downstream of the front, much like the lowering of the water level in subcritical flow over a hump in the bed.

A simple analysis of the hydraulics of the situation along these lines suggests that the front will be engulfed when

$$\frac{V}{\sqrt{gt\,(1 - s_i)}} \leq \sqrt{2}\,\left(1 - \frac{t}{h}\right)$$

where V is the average approach velocity, h the approach depth, t the accumulation thickness, and s_i the relative density of the ice (Pariset, Hausser and Gagnon, 1966). A rearranged version of this relation is compared with experimental and field results in Figure 11. The agreement is noteable. Figure 11 indicates that an ice accumulation cannot progress upstream if $F = V/\sqrt{gh} > 0.16$ and that the dimensionless thickness, t/h, of the front portion of the accumulation must be less than 0.33. Field measurements (Kivisild, 1959) suggest that the critical value of F is actually about 0.08-0.10.

If the approach velocity is such that $F > 0.1$ (e.g. about 1.1 m/s for 10 m depth) no accumulation is possible (i.e. any accumulation will be continually engulfed) until a backwater due to some obstruction downstream reduces the velocity below the maximum accumulation velocity. If this occurs the accumulation should then progress, leaving behind a thickness that, initially, is close to 1/3 of the flow depth.

Such are the hydraulic restraints placed on a floating ice jam.

Structural Constraints

As an accumulation progresses upstream an increased area of accumulation is exposed to the drag of the flow passing underneath. This accumulated drag must be transferred back to the original obstruction, or to the banks of the stream. To transfer this load, the accumulation must be strong enough to sustain it. As pointed out by Pariset et al. (1966) the compressive strength of the accumulation is a direct function of its thickness. If the thickness given by the hydro-dynamic constraints is insufficient to sustain the load to be transferred, the accumulation will collapse or shove until it is thick enough. The channel is considered narrow if the maximum thickness of the accumulation given by the hydraulic constraints is sufficient to sustain the additional load from an advance of the front by shear on the banks. In this case the accumulation thickness is governed by the hydraulic constraints.

The channel is considered wide if the accumulation shoves as it lengthens. The shoving increases the maximum thickness until the drag added by an advance of the front is sustained by shear on the bank as shown in Figure 12. When this thickness is reached no additional load is transferred to the obstruction. Thereafter the maximum accumulation thickness remains the same despite a lengthening of the accumulation. The thickness left behind as the accumulation advances is then governed by the strength of the accumulation - that is, by the structural restraint. This maximum thickness has come to be called the equilibrium thickness.

The strength of an accumulation of ice and therefore, in a wide river, its thickness depends on parameters μ, which is related to the porosity and internal friction of the accumulation, and C_i, a cohesion parameter. The maximum accumulation thickness is given by (Michel, 1965; Pariset et al., 1966; Uzuner and Kennedy, 1976).

$$\mu\rho_i\ (1 - s_i)gt^2 - [(g\rho_i S - \frac{2C_i}{B})B]\ t - \tau\ B\ =\ 0$$

where S is the channel slope, B the channel width at the bottom of the accumulation, and τ is the shear of the water on the bottom of the accumulation as shown in Figure 13.

Similar values of μ have been determined from field measurements in at least two independent investigations (Pariset et al., 1966; Beltaos, 1978) and these values are not inconsistent with those found in the laboratory (Uzuner and Kennedy, 1976). From these investigations a value for μ of 1.2 seems reasonable. Little is known about the cohesion parameter, except that its effect seems to be small in breakup jams. Laboratory tests suggest $C_i \simeq 100 - 500$ Pa.

For uniform flow under the accumulation

$$\tau\ =\ \rho g h_i S$$

in which h_i is the distance from the bottom of the ice accumulation to the maximum velocity point. The ratio h_i/h_j, where h_j is the depth of flow beneath the jam, can be found from given roughnesses by

$$\frac{h_i}{h_j}\ =\ \frac{1}{1 + k_r^{-1/4}}\quad \text{where } k_r\ =\ \frac{k_i}{k_b}$$

In turn, h_j can be found from

$$\frac{V_j}{V_*} = 2.5 \ln \frac{R}{k} + 6.2$$

where $V_j = \frac{Q}{h_j B}$: $R = \frac{h_j}{2}$: $k = k_b \left(\frac{1 + k_r^{1/4}}{2}\right)^4$: $V_* = \sqrt{gRS}$

The roughness of the ice cover seems to be related to the thickness of the ice cover (Kennedy, 1958; Nezhikhovskiy, 1964; Tatinclaux and Cheng, 1978) and, reason would suggest, the size of the ice floes. That is

$$\frac{k_i}{\ell_i} = f\left(\frac{t}{\ell_i}\right)$$

where ℓ_i is a typical floe length. A crude estimate of the form of this function is shown in Figure 14.

To calculate the equilibrium accumulation thickness all these relations must be satisfied simultaneously. A suggested procedure is

1. Estimate k_i
2. Calculate h_j
3. Determine B
4. Calculate h_i
5. Calculate τ
6. Calculate t
7. Calculate water level = $h_j + 0.9t$

A typical calculation is detailed in Appendix I. Consideration of the above will indicate that the increased water level is caused both by the additional roughness, and the additional thickness of the accumulation, over that of the normal solid ice cover.* On large flat rivers the former is the more important influence. In smaller steep streams the latter would probably be more important.

Because it is based on uniform flow calculations the above calculations give an estimate for the maximum level along a floating ice jam. However the actual water level will follow a gradually varied flow profile as sketched in Figure 7(a). An actual example is shown in Figure 7(c). The calculation of such profiles is not considered herein, but are little more complicated than gradually varied flow calculations for normal open channel situations if the downstream boundary condition can be determined. This however is difficult at present.

It is important to keep this open channel behaviour of ice covered channels in mind when assessing water levels along such channels.

* If the ln expression for velocity is replaced by a power function approximation, steps 4-7 collapse into the evaluation of the single expression given in Appendix 1. If, further, B varies little with h_j, over the values of h_j of interest, this would include steps 2 and 3 too. The siple relation for maximum ice jam stage that then results is given in Appendix 1 and shown in Figure 15.

In all the above investigations the possibility of channel bed changes have not been considered. Yet such changes could have an important influence on the behaviour of the jam and the water levels it causes, and presumably on such engineering structures as buried pipelines, bridge piers or spur dykes that lie on or under the bed. A field observation of such scour is discussed below. A first attempt to calculate scour under a quasi-steady floating ice jam has been reported by Mercer and Cooper (1977).

Stability of Floating Ice Jams

If the situation that prevailed at formation changes, the accumulation configuration may change. For example if the discharge increases the accumulation can be expected to shove and thicken. However, if the discharge is reduced little should change, other than the water levels. Andres (1980) took advantage of this in their analysis of the 1978 ice jam at Fort McMurray. Likewise if the jam is thickened by the deposition of ice entrained upstream it should simply increase the upstream water levels. On the other hand, if the accumulation begins to melt it can become thin enough to be unstable and shove again.

GROUNDED ICE JAMS

These jams can be caused, for example, by the collapse of a floating ice jam, the sudden stoppage of a surge of ice and water, or by blockage of the flow under a hanging dam. Given the limited and irregular depths of most natural channels, the formation of such jams is an obvious possibility. The destructive jams described by Barnes (1928) and Frankenstein and Assur (1972), on the Allegheny and Israel Rivers respectively, were known to be grounded. The description of the Moira River ice jams at Belleville (Lathem, 1974) suggests that they only became threatening when the passage under the ice was blocked - that is, when the jam primed. Mathieu and Michel (1967) found that if the ratio of the flow depth beneath a floating jam was less than the largest dimension of the entrained floes, the jam would 'prime' and become a grounded jam.

As stated by Michel (1971) "in such jams the headlosses are considerable compared to those of a simple [floating] jam. It has been impossible to determine these losses in a general manner because of the seemingly fortuitous length of grounding in each case and the variable solidarity of the accumulation of the floes". This states the problem succinctly. However, given the possibility that such jams may be responsible for the highest breakup water levels, much further work is required on this type of jam, if only to establish a reasonable upper limit on the high water levels possible.

ICE JAM FORMATION AND FAILURE: THE UNSTEADY CASE

Almost all past work has been concerned with almost-steady flow past an ice jam. However an examination of reports recorded in archives and told by eye-witnesses reveals important features of observed ice jams that are difficult to explain from steady flow considerations. A minor but typical example is provided by Johnson and Kist ner (1967). During breakup of the Meade River on the north slope of Alaska "a flow of brownish river water about 40 cm in height was progressing over the top of the river ice (June 7) ... at the pace of a fast walk, perhaps 8 km/h. A floe [sic] of jumbled ice blocks choked the channel behind the slush wave. This ice flow [sic] at times overflowed the unbroken ice or simply created ice blocks as it advanced. The advancing ice flow [sic] with its slurry of water and ice blocks jammed quite suddenly when it reached a narrowing of the channel 0.5 km below camp. The river, now completely choked with jumbled ice blocks, rose rapidly, about 2 m in 1.5 hours ....

On the afternoon of the 10th a very high water level allowed the ice jam to slip downstream ... evidently a similar ice jam had broken upstream ... this time considerable ice arrived from upstream and the river was choked with ice blocks for several kilometres upstream [see Figure 16a]. On the night of the 11th the entire ice floe [sic] broke .... After the dam [sic] released, the river level dropped briefly on the 11th and again on the 13th leaving both banks lined with vertical cliffs of ice blocks 3-4 m high (Figure 16b).

A characteristic of the more dramatic reports is the extremely rapid rise in water levels. For example, in the Athabasca River at Fort McMurray in 1875 "in less than an hour the water rose 57 feet, flooding the whole flat and mowing down trees, some 3 ft. diameter, like grass ..." (Moberly and Cameron, 1929); on the Peace River near the Mikkwa River confluence in 1886 "the ice in the Peace River struck during the night and about 2 a.m. the water rose rapidly in the Red [Mikkwa] River. Two feet more of rise would have put it over the banks ..." (Hudson's Bay Co. Journal, Red River, 1886); on the Athabasca River 35 km upstream of the House River confluence in 1936 "During the night they [three men] awakened to find three feet of water in the room. Scrambling into some clothes they waded out and untied their horses and tried to find higher ground. The water rose so rapidly that all they could do was to climb a tree. Lee and Cinnamon got a safe one and climbed higher as the water rose. They could see Donaldson in difficulties and shouted to him, but he appeared unable to climb or the sapling would not support him and he gradually sank out of sight ..." (Athabasca Echo, 27 April 1936, Athabasca, Alberta); on the Red Deer River near Red Deer the water rose 11 m in about 3 hours and removed the superstructure of a CNR bridge (Morris, 1976).

Such rapid increases can only be explained by the action of surges created by the failure, and perhaps the reformation, of ice jams. That such surges occur is supported by the several reports in the literature of very high velocities. For example Killaly (1887) observed "the ice [on the Missouri River] in the neighborhood of St. Joseph ... came down from above with a rush, causing a sudden rise in the river .... The river foamed and hissed. The whole waterway was filled with broken ice grinding along the bottom, and pitching and tossing on the surface. The water itself was not to be seen, as the mass of broken ice, and drift rolled by - forest trees and masses of brush, wreckage of all sorts, whirling around, and forced into the air by the upward action of heaving ice. A gorge [jam] had broken above ...." Doyle (1977) reports on breakup in 1977 on the Athabasca River at Fort McMurray: "Flood wave estimated to be 5 m high rushes downstream past bridge tossing ice blocks into air as it passes at an estimated velocity of 5 - 6 m/s". With such behaviour possibly preceding the formation of an ice jam it is difficult to imagine they would take up the orderly characteristics envisaged when analysing steady, floating jams. In particular, the increased possibility of priming a grounded ice jam when such 'ice surges' are halted to reform a jam is obvious.

Consideration of the result of a sudden halting of such a surge suggests the answer to another anomaly. The quotation given above reports a 17 m increase in water level just after the passage of a surge on the Athabasca River at Fort McMurray in 1875. If this is simply caused by the passage of a surge released by an ice jam failure upstream, this ice jam would have had to be at least double this height - say about 35 m high. Although such an ice jam may be possible in the deep valley of the Athabasca River upstream of Fort McMurray, it is unlikely. However, if the consequence of a surge reflection caused by the sudden reformation of the jam downstream of Fort McMurray is considered, a much lower initial surge, and hence a lower upstream ice jam, is required to explain the increase in water level noted.

This line of reasoning, and the analysis of surges created by ice jam formation and failure has been pursued by Henderson and Gerard (1981). This analysis considered the consequences of sudden complete failure and subsequent reformation of ice jams. It confirmed the change in water level downstream of an ice jam immediately after failure cannot be more than half the initial water level difference across the jam. It also showed that extremely high velocities can be expected downstream of such a failure. A field example of high velocities after a partial jam failure has been reported by Gerard (1975). The 2-3 m standing waves created by this sudden discharge is shown in Figure 17. Figure 18 shows another example on the Yellowstone River in Montana. Doyle (1977) reports velocities as high as 6 m/s caused within an ice jam as it readjusted within an ice jam as it readjusted. Both Henderson and Gerard (1981) and Beltaos and Krishnappan (1981), the latter using numerical techniques, have investigated the behaviour of the jam documented by Doyle (1977) and report good agreement between prediction and observation. Measurements of the propagation of surges, both in the upstream and downstream directions, have been reported by Calkins (1981). Although often of short duration (from minutes to hours) the possibility of unusual scour by such events is obvious; to quote Killaly (1887) again "On the 24th [February] a gorge occurred .... The river hurled itself, with great force, against dyke No. 6, and washed along its face ... in a few hours the whole face of the dyke had been undermined; the channel having scoured out a depth of thirty-four feet [from the account this seems to represent about 4 m of scour]. The dyke 'turned over'!"

BREAKUP WATER LEVELS

As mentioned before, a major incentive for developing an understanding of ice jam behaviour is the need to predict breakup water levels for river engineering design purposes. These are often more important than water levels caused by summer, or open water, floods. They should therefore be subject to at least as much scrutiny in a river engineering investigation.

Analytical Estimates

Some indication of what these levels might be can be determined by analysis.

Lower Bound

If no ice jams are expected to form at the location of interest the breakup water level will be closely related to the freeze-up water level. As discussed previously indications are that, for a reasonably competent floating ice cover, breakup will occur when the water level rises about a metre or so above the maximum winter stage. This relation can be refined for a particular site if some observations on the time of breakup are available.

After the relationship has been established breakup water levels for various past years can be estimated from winter discharge records and estimates of the thickness and roughness of the ice cover at the time of maximum winter stage. A probability analysis can be carried out on these estimates to fix a lower bound on the breakup stage distribution. It should be noted that in many locations these no-ice-jam levels will be above the 2-5 year summer flood levels.

Upper Bound

On the assumption that only floating jams can form and that they form downstream of the site each year, an upper bound for the probability distribution of breakup water levels can be estimated using discharge records and the analysis of floating ice jams described above.

If no grounded jams form the actual probability distribution should be somewhere between these bounds, depending on the probability of an ice jam forming in the reach each year. Unfortunately, this probability is difficult to determine. The other limitation on the above analysis is that jams other than simple floating jams may form in or near the reach of interest. As pointed out above, the present understanding of breakup events other than quazi-steady floating jams is very poor.

Hence because of these limitations on the current ice jam state-of-the-art, the above deterministic estimates must be supplemented by as much information on actual past breakup water levels as possible.

Empirical Estimates

As noted previously breakup water levels are very site-specific. Therefore to be useful the water level records must come from very near the site of interest. Sometimes information is available from residents, whether permanent or itinerant (eg. farmers, trappers). Other times information can be gleened from archives of a nearby community (newspapers, biographies, maintenance records, journals, family photographs, etc.). In some cases a standard hydrometric gauge is installed in or near the reach, although failure of these installations during breakup is common. If such a gauge exists the original chart recordings or field notes must be examined. If an ice jam did form the water level changes may be rapid and will make interpretation of the chart difficult. An example is shown in Figure 19.

However, more often than not, there are neither inhabitants nor gauges near the reach of interest. The only available information is then that which can be deduced from environmental evidence such as trim lines, windrows, and damaged vegetation. Of the latter the most important items are the ice scars left on trees by high ice, an example of which is shown in Figure 20. The elevations of these scars provide a lower bound on the higher breakup water levels that have occurred during the life of the trees. If the scars are sampled as shown in Figure 21, and their age determined by tree-ring dating (Sigafoos, 1964; Parker and Lozsa, 1973), an approximate history of past high breakup water levels can be reconstructed. A typical record completed in this way is shown in Figure 22.

On the basis of this observational data, both historical and environmental, another estimate of the breakup water level probability distribution can be made. A method for carrying out a probability analysis of such unorthodox data is described by Gerard and Karpuk (1979), exerpts of which are included herein as Appendix II.

An engineering assessment of the results of the analytical and empirical investigations will allow a compromise probability distribution for breakup water levels to be chosen. This should then be combined with the estimated probability distribution for summer floods to get the required probability distribution for design.

Joint Probability Analysis

The two types of floods are more or less independent and are not mutually exclusive (ie. both can occur in a given year). Hence the probability of one or both exceeding a given stage in a year, P, is given by

$$P = P_b + P_s - P_b P_s$$

where P_b = probability of a breakup flood exceeding the chosen stage in a year;

P_s = likewise for summer floods.

This joint probability will obviously be higher than either of the other two. A typical situation is shown in Figure 1(a).

Maximum 'Probable' Breakup Water Levels

As for summer floods it is very useful to have some estimate of the maximum breakup water level that could occur. Like all things associated with ice jams, this is difficult to assess. The potential is exemplified by the following description of an ice jam on the Yukon River (Henry, 1965): "The highest jam causing the greatest depth of flooding, according to reliable reports, occurred at Ruby, Alaska. Ruby is built on a hillside, one of the few villages situated well above the river. In the spring of 1930 a big ice jam formed and the water backed up to the porch level of the present Northern Commercial Company store. Boats, tied to the porch, were at least 35 feet above normal river levels. The river valley is 12 to 15 miles wide at Ruby and remains about the same for miles downstream. So the jam extended at least 15 miles across [sic] and rose to a height of 65 feet. No one knew the location of the blocking jam down river."

With a long well-gounded jam in an entrenched valley the water level is presumably limited only by the discharge and the supply of ice from upstream - the latter being a constraint that should not be overlooked. However, in a reach with a well-developed flood plain, water will be able to move around the toe if the water level rises above the flood plain. The maximum water level should then be a metre or so above the lowest passage on the flood plain. This mechanism limited the water level of the 1963 ice jam on the McLeod River in Alberta shown in Figure 23. (Note that levee construction to provide protection against summer floods could remove this safety valve.)

Although a particular reach may be free of grounded jams it may still be within the backwater from a grounded jam in an entrenched reach downstream, or in the path of a surge released by the sudden failure of one upstream.

Hence at present little more than a qualitative assessment of maximum breakup water levels is possible, but nevertheless such an assessment should be made.

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