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RIVER ICE MANUAL
CHAPTER 5. PREDICTIVE METHODS

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ABSTRACT

This report summarizes in simple form, the predictive capabilities pertaining to the regime of ice in rivers. Starting with the cooling of water in the fall, basic relationships are presented describing the formation of ice, its transport and eventual accumulation into ice covers, thermal ice growth in the winter and decay in the spring. Break-up processes and recent developments in forecasting are described next. Ice jamming during breakup, including surges from ice jam release is the final topic.

RÉSUMÉ

Le présent rapport résume, sous forme simple, les capacités de prévision dans le domaine du régime des glaces de cours d'eau. Les données fondamentales sur la glace de rivière, d'abord le refroidissement de l'eau à l'automne, puis la formation de la glace, son transport et son accumulation éventuelle sous forme de couvertures de glace, la croissance thermique de la glace en hiver et la décroissance au printemps sont décrites ici. Les auteurs traitent ensuite des débâcles et des récents progrès dans le domaine de la prévision, et finalement, de la formation d'embâcles au cours du dégel, y compris les crues causées par la rupture des embâcles.

MANAGEMENT PERSPECTIVE

In 1987, the New Brunswick Subcommittee on River Ice initiated the production of a River Ice Manual, intended for non-specialists ranging from the general public to engineers and water managers. Emphasis is on processes that are particularly relevant to New Brunswick, a province where river ice is a major cause of flood and related damages.

This report is intended to become Chapter 5 of the Manual and summarizes the basic quantitative river ice knowledge for the entire season, that is, fall and water cooling, freeze up and ice cover formation, winter and ice growth, spring thaw, ice decay and breakup with the attendant jamming and flooding potential.

PERSPECTIVE-GESTION

En 1987, le sous-comité sur la glace de rivière du Nouveau-Brunswick a entrepris la production d'un guide de la glace de rivière à l'usage des non-spécialistes, du grand public, des ingénieurs et des gestionnaires de l'eau. L'accent est mis sur les processus particulièrement importants pour le Nouveau-Brunswick, province où la glace de rivière est l'une des principales causes d'inondations et d'autres dommages.

Le présent rapport deviendra le chapitre 5 du Guide; les auteurs résument les renseignements quantitatifs de base sur la glace de rivière pour toute l'année, c'est-à-dire le refroidissement de l'eau à l'automne, le gel et la formation d'une couverture de glace, la croissance de glace en hiver, le dégel printanier, la décroissance de la glace et la débâcle ainsi que les problèmes potentiels d'inondations et de formation d'embâcles qui y sont associés.

FREEZE UP

Water Cooling

Heat exchange at the open water surface is the predominant process by which river temperature drops in the fall. Heat transfer takes place due to: solar or short-wave radiation; long-wave radiation; evaporation or condensation; convection; and precipitation. Corresponding heat fluxes (=amounts of heat transferred per unit area and per unit time) can be calculated according to hydrothermal and meteorologic principles (Tsang, 1982; Ashton, 1986). Minor heat exchanges may also occur at the stream bed, due to groundwater flow; heat stored in bottom sediments; geothermal flux and flow friction. These are generally negligible but may become significant when an ice cover is present. Figure 1 shows how three major components of heat loss rates vary with temperature difference between water and air and with wind speed, for an assumed set of meteorologic conditions. Table 1 illustrates how the total heat loss rate varies with temperature, wind speed and cloud cover.

For simplicity, the total or net heat flux, ϕ^*, is often expressed as

$$\phi^* = K\,(T_a - T_w) \qquad (1)$$

in which T_a, T_w = air and water temperatures; and K = empirical heat exchange coefficient that accounts for local conditions and meteorological factors. Values of K between 20 and 60 $W/m^2{}^\circ C$ have been found for the St. Lawrence River (Shen et al, 1984; Prowse, 1987).

Knowing ϕ^* and stream hydraulics makes it possible to predict T_w as a function of river location and time, usually by computerized solution of the differential equation expressing conservation of heat. Ice formation is imminent when T_w drops to near 0°C.

Border Ice

The first ice to appear on a river usually forms along the banks where flow speed is low. Border ice grows vertically and laterally toward the mid-stream. Lateral growth can take place even with slightly positive mainstream temperatures, depending on flow velocity and meteorologic conditions (see, for example, Fig. 2). However, the rate of lateral growth can only be predicted empirically, based on observation (e.g. Newbury, 1968).

Moving Sheet Ice

Under certain conditions, a thin layer of solid ice may form on the water surface without being attached to the shores. Matousek (1984) reasoned that this phenomenon is due to the stream turbulence being unable to overcome the rise velocity of ice crystals that form, and thus remain, on the surface. His analysis indicated that there is a "critical" average flow velocity below which moving sheets form; this velocity depends on heat transfer parameters and hydraulic resistance (Fig. 3). An empirical equation has also been proposed by Marcotte (1986). Ice sheets formed in this manner are very thin (maximum observed thickness = 30 mm) but can attain very large horizontal dimensions, comparable to the channel width. Once halted, these sheets cause rapid freezeover of the entire river surface.

Frazil Ice

In most instances and over the main part of river width, a slight supercooling of the water (to -0.1°C) results in formation of tiny ice particles called frazil. The amount of frazil ice produced from an open-water area, A_o, is given by:

$$Q_i = \frac{\phi^* A_o}{\rho_i L} \qquad (2)$$

in which Q_i = volume of ice produced per unit time; and ρ_i,L = ice density and latent heat of fusion, respectively. Exactly how frazil forms is still a matter of debate (Ashton, 1986). Williams (1972) studied frazil formation in the Ottawa River and found that large amounts were formed at this study site when the total heat loss rate exceeded 360 W/m^2.

In supercooled water, frazil particles are "active", i.e. they stick to each other or to any submerged surface. Particle collisions lead to formation of frazil "clusters" or "flocs" whose larger size makes them more buoyant so that they quickly rise to the surface. There, they agglomerate, eventually forming rounded floes called ice "pans" or "pancakes". Typically, such floes consist of a top layer of solid ice (a few centimetres thick) and a much thicker accumulation of porous slush underneath.

Anchor Ice

In highly turbulent reaches, supercooled water or active frazil can reach the river bed where they form an accumulation called "anchor" ice. Little is known about anchor ice formation, growth and detachment, despite its potentially large effect on hydraulic resistance or ice discharge. A first step in predicting anchor ice processes was made recently by Marcotte and Robert (1986) who developed an empirical mathematical model.

Icings

Another form of river ice that usually occurs in arctic or subarctic shallow streams, is the icing or "aufeis" or "naled". It grows by repeated freezing of thin layers of water flowing over existing ice. Icings of immense dimensions have been observed (Ashton, 1986; Prowse, 1987) and serious associated problems have been reported, e.g. flooding, damage to structures, blockage of culverts and water intakes. Prediction of aufeis development is uncertain (Ashton, 1986). Schohl and Ettema (1986) performed laboratory experiments and formulated fundamental dimensionless parameters of the icing process.

FREEZE UP JAMMING AND COVER FORMATION

Freeze Up Jams

Full-width ice cover can start by gradual extension and closure of border ice; arrest of moving sheet ice; and congestion of moving ice pans. The latter process is the most common but not reliably predictable without field observations.

Ice floes that come to a halt may submerge and continue moving downstream if the local velocity exceeds a critical value, V_S. Early field studies indicated that $V_S = 0.7$ m/s for frazil and slush (Joint Board of Engineers, 1927). More recent work suggests that V_S depends on floe thickness, porosity, shape and horizontal dimensions (e.g. see reviews by Ashton, 1986; Beltaos, 1986).

Where congestion occurs and the flow velocity, V, is less than V_S, a surface jam, i.e. a loose cover comprising a single layer of ice floes, is initiated. This cover solidifies by freezing of the interstitial water. Suppose next that the leading edge of this type of cover arrives at an area where $V > V_S$. The incoming floes will submerge and two types of jams may then form, depending on local conditions.

(i) Submergence and deposition - frontal progression of cover. The flow velocity is low enough to cause deposition of submerging blocks at the front. A thickened, porous cover will then form and progress upstream with a thickness, t (Michel, 1971)

$$t = V_u^2 / 2(1 - S_i)(1 - p)\,g \qquad (3)$$

in which V_u = average flow velocity under the accumulation; S_i = specific gravity of ice, 0.92; g = acceleration due to gravity and p = porosity of the accumulation, 0.4-0.9 depending on degree of consolidation of the cover. This type of cover is also known as a "narrow" jam (Pariset et al., 1966).

(ii) Submergence, transport and eventual deposition - hanging dam. Submerging blocks are transported by the flow until they come to a region of reduced velocity and are thus able to deposit under the cover. Very large accumulations, called hanging dams, can form in this manner. Michel (1984) recommended the range 0.6 to 1.3 m/s for the threshold transport velocity of ice floes, based on empirical evidence (see also lab studies by Taticlaux and Gogus, 1981).

As a porous ice accumulation, or jam, propagates upstream, the forces applied on it increase and, under certain conditions, may exceed the jam's capacity to resist them. The jam collapses or "shoves" and thickens until it is just able to resist the applied forces. Because this phenomenon is much more frequent during breakup, it will be discussed in more detail later, in the "Breakup" section.

Solid Ice Cover

Once a stable, porous cover has formed, freezing of intersticial water solidifies a surface layer which grows downward by freezing at the ice-water interface. Heat loss is retarded by the ice cover itself and by snow that may be present. Where freezing takes place into a frazil accumulation, the rate of thickening of the solid ice layer is augmented by the factor 1/p (Calkins, 1979). A simple, semi-empirical formula is often used to calculate the solid ice thickness

$$h_i \text{ (in cm)} = a_1 \sqrt{D_F} \tag{4}$$

in which D_F = accumulated degree-days of frost (°C-days) and a_1 = empirical coefficient evaluated by calibration. Where no local data exist, a_1 can be estimated as follows: a_1 = 2.7 for windy lake without snow; 1.7-2.4 for average lake with snow; 1.4 - 1.7 for average river with snow; and 0.7 - 1.4 for sheltered small river with rapid flow. In New Brunswick, several agencies record river ice thicknesses and their data may be useful in a variety of applications (see summary report by LeBrun-Salonen, 1983). For the St. John River at Fredericton, Bray and Boyer (1977) determined a value of 1.9 for a_1.

BREAKUP

The decay, fracture, transport and eventual clearance of the ice from a river, are the main processes taking place during the so-called breakup period. Breakup is triggered by mild weather and is of particular interest because it is attended by major ice jams of serious potential for damage.

Ice Decay

Mild weather brings about reductions in ice thickness and strength. Melting of the cover can occur at both top and bottom surfaces and is a complex hydrometeorologic process. Bilello (1980) proposed a simple formula for thickness reductions due to mild weather:

$$\Delta h_i = a_2 D_T \tag{5}$$

in which D_T = accumulated degree-days of "thaw" (above a base of -5°C for rivers); and a_2 = empirical coefficient, between 0.4 and 1.0 cm/°C-d for N. Canadian and Alaskan rivers.

Ice strength is reduced by penetrating solar radiation, following reduction of ice temperature to 0°C. Excess heat melts the ice at crystal boundaries (Bulatov, 1972; Ashton, 1983). In extreme cases, the result is the well-known "candled" ice whose strength is practically nil. Prediction of ice strength reduction is, however, complex and uncertain where no calibration data exist (e.g. see Prowse, 1987).

Ice Fracture

An ice cover that retains some of its strength may be fractured in different ways, i.e.:

- Hinge cracks: these are longitudinal fractures located near the shores and caused by uplift pressures that develop due to increasing discharge (Beltaos, 1985; Billfalk, 1981). In narrow streams only a single crack develops.
- Transverse cracks: these are lateral fractures, spaced a few river widths apart and likely caused by horizontal bending (Beltaos, 1985). The cover thus becomes a sequence of separate ice sheets.
- Impact breaking: where large ice sheets are able to move in the river, impacts between sheets or against channel boundaries cause rapid breakdown to block-sized fragments.
- Breaking front: the release of major jams causes very steep flood waves which, in turn, may effect in-place fragmentation of downstream covers. Breaking fronts are known to occasionally move at high speeds (e.g. 5 m/s) for long distances (Prowse, 1987) but the breaking mechanism is not clear. Ferrick et al. (1986) presented novel field data on this phenomenon, along with insights as to its causes.

Initiation of Breakup

When a runof event is forecast for a specific river reach, it is important to be able to predict whether it will cause breakup of the ice cover. To tackle this problem in a quantitative manner, it is necessary to define what is the "onset" of breakup or the breakup "initiation". The prevailing definition pertains to the time when the ice cover at a given site is set in motion for a sustained period. This definition is "tied" to the jamming that occurs shortly afterwards and is meaningful in all instances but where the cover is destroyed by a breaking front or disintegrates in place by thermal inputs.

Using the above definition, Beltaos (1984a) formulated a breakup initiation criterion, based on the premise that movement of the ice cover begins when the ice sheets formed by transverse cracking are able to negotiate bends or other obstacles. This is made possible by increasing river stage and water surface width. Beltaos' criterion requires detailed information on river geometry and ice cover width. A simpler empirical approach has been used in the past whereby the breakup is expected to start when the water level rises above that of the preceding freeze up by a "critical" amount, ΔH_B. In turn, this rise depends on ice cover thickness, h_i, and strength (see also Beltaos, 1984b; and Shulyakovskii, 1963), i.e.,

$$\Delta H_B = C\, h_i \tag{6}$$

in which C = site-specific dimensionless coefficient. Beltaos (1989) showed that his breakup initiation concept is consistent with Eq. 6 while C is dependent on several variables such as river curvature and slope, flow shear stress, ice strength and thickness loss, and steepness of the river banks. Where the thermal effects (thickness/strength losses) on the ice cover are small ("premature" breakup), the value of C has a maximum of C_o at a given site. Table 2 summarizes values of C_o determined from data at six Canadian river sites where they generally fall in the narrow range of 2.2-3.5. The large C_o for the Thames River is due to that stream's low water surface slope and steep banks. Where the thermal effect is significant, the difference, $Ch_i - \Delta H_B$, has been empirically related to a thermal index (of which the simplest version is the accumulated degree-days of thaw). Such relationships are site-specific (e.g., see Beltaos 1984b; Tang and Davar, 1984; Tang et al., 1986) and more field data are needed to enable generalization.

BREAKUP JAMS

Once breakup has started, ice jams form and largely control subsequent developments. There is little one can do about predicting where and when jams will form, other than refer to past experience. Breakup jams are held in place by sections of intact ice cover and can form anywhere in a given reach. Nevertheless, there are preferred sites of formation, depending on the presence of jam-conducive geomorphic or man-made features (e.g. sharp bends, bridge piers, shallows, slope reductions, etc.).

Assuming or knowing that a jam has formed somewhere in a river, it is possible to predict the water levels that the jam could cause. Pariset et al. (1966) distinguished between "narrow" and "wide" channel jams, depending on their ability to resist the applied forces; these increase with stream width. The "narrow" type is stable with the thickness given by Eq. 3 while the "wide" jam forms after the collapse of a "narrow" one, as discussed earlier. As a rule, breakup jams are the least capable to resist applied forces and tend to be of the wide kind (Beltaos, 1983), thus having greater potential for flooding and damage, other things being equal. However, flow discharge at breakup is usually much greater than that at freeze up which further contributes to the flooding potential of breakup jams.

Calculation of actual stages caused by ice jams can be so complex as to require numerical applications (e.g. see Beltaos, 1986; Beltaos and Wong, 1986; Petryk et al., 1981). On the other hand, for quick estimates of potential jam levels, one could consider the "equilibrium" condition which defines the largest water depth attainable (see Fig. 4). This depth can be estimated very simply by means of Fig. 5 or, in a more detailed manner that accounts for channel geometry and jam roughness, by the method described by Beltaos (1983) or by MacLaren Plansearch (1985).

When a jam suddenly lets go, a steep water wave moves downriver in surge-like fashion and can cause serious damage with little warning. The conditions leading to jam release are very little understood which translates to uncertainty in either forecasting such events or inducing them as a flood relief measure. It is possible, however, to roughly predict the surge characteristics via hydrodynamic principles. Ordinarily, this requires computer use (Mercer and Cooper, 1977; Beltaos and Krishnappan, 1982). Crude estimates can be made via the simple theory of Henderson and Gerard (1981), summarized in Fig. 6. It may be noted that the severity of the surge increases with the ratio H_J/H_D that, in turn, can be shown to increase with stream size. Large rivers should therefore experience more severe surges which is in accord with experience.

As an example of ice jam predictions, consider a 600 m-wide river, with a slope of 0.3 m/km. Let the discharge per unit width be 3.0 m^2/s (i.e. total discharge = 1800 m^3/s). To find the potential stage of a jam formed under these conditions, we may use Fig. 5. The parameter ξ works out to be 81 whereby η = 55 and H_J = 9.9 m, a considerable depth that, on some rivers, might translate to flooding. To get an idea of what might occur downstream upon release of this jam, we may use Fig. 6, assuming that H_D = 4.1 m, a value either given or estimated. We first calculate F_D and H_J/H_D as 0.12 and 2.4, respectively. With these values, Fig. 6 gives $(H_S - H_D)/(H_J-H_D) = 0.41$ which implies H_S = 6.5 m. It is difficult to use Fig. 6 for C and V_S because interpolation would be uncertain. We could, however, utilize the following two equations derived from the original paper, i.e.:

$$\frac{C}{\sqrt{gH_D}} = F_D + \sqrt{0.5 \left(\frac{H_s}{H_D}\right)\left(\frac{H_s}{H_D} + 1\right)} \tag{7}$$

$$\frac{V}{\sqrt{gH_D}} = F_D + \left(1 - \frac{H_D}{H_S}\right) \sqrt{0.5 \left(\frac{H_s}{H_D}\right)\left(\frac{H_s}{H_D} + 1\right)} \tag{8}$$

Since H_S/H_D = 6.5/4.1=1.58 we obtain C = 9.8 m/s and V = 4.1 m/s. The latter is an unusually large water velocity that under open-water conditions may occur only during extreme floods. The celerity, C, of the wave is also very large and would cause very rapid rise of the downstream water levels. By contrast, in a 100 m-wide stream of identical slope and discharge per unit width, we find H_J = 5.2 m, H_S = 4.6 m, C = 7.7 m/s; and V = 1.5 m/s.

APPLICATIONS

Practical questions pertaining to river ice usually involve such matters as forecasting and flood warning, selection of remedial measures and impact assessment of structures altering the hydrologic regime. It is possible in some instances to study such questions by simple applications of the quantitative understanding and formulae summarized in this section.

Often, however, it is essential to simulate in detail the entire ice regime, e.g. freeze up to breakup, or a portion of it, e.g. breakup. This can be accomplished using numerical, computer assisted, models that are based on equations such as the ones presented herein while, in addition, being capable of taking into account the detailed geometry of the river and weather variations. As has already been pointed out, however, existing river ice knowledge has serious gaps. This is commonly circumvented by making "plausible" empirical assumptions backed by site-specific observation.

Occasionally, the nature of the problem is such that little faith can be placed on existing data or mathematical analysis. Physical modelling might then be an alternative or complementary approach. A major difficulty here lies in the scaling down of the strength characteristics of intact ice covers when their behaviour is relevant to the problem at hand (e.g. see Michel, 1978; Wong et al., 1988).

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Table 1. Total heat loss rates at water surface (from Asvall, 1972). Constructed from measurements in Norway.

Cloud Coverage 0 - 10	Wind Speed m/s	Heat Loss Rate (W/m^2)			
		Air Temperature (°C)			
		0	-10	-20	-30
0	1	138	280	402	528
	5	147	390	600	785
5	1	80	222	343	470
	5	88	336	541	726
10	1	26	164	288	414
	5	34	277	486	671

Table 2. C_0-values at six Canadian river sites.

Site and Latitude (N)	Source	Long-term Mean Discharge (m^3/s)	Water Surface Slope (m/km)	Average Ice Thickness before Breakup (cm)	C_o
Thames River at Thamesville 42°32'42"	Beltaos (1989)	51	0.23	21	8.0
Grand River near Marsville 43°51'43"	Beltaos (1989)	7.7	2.3	35	2.2
Ganaraska River near Dale 43°59'07"	Beltaos (1989)	3.4	1.8	29	3.5
Nashwaak River at Durham Bridge 46°07'33"	Beltaos (1989)	36	0.73	60	2.5
Meduxnekeag River at Belleville 46°12'58"	Tang *et al.* (1986)	26	1.8	51	3.1
Moose River at Moose River 50°48'50"	Beltaos (1989)	780	0.38	69	2.8

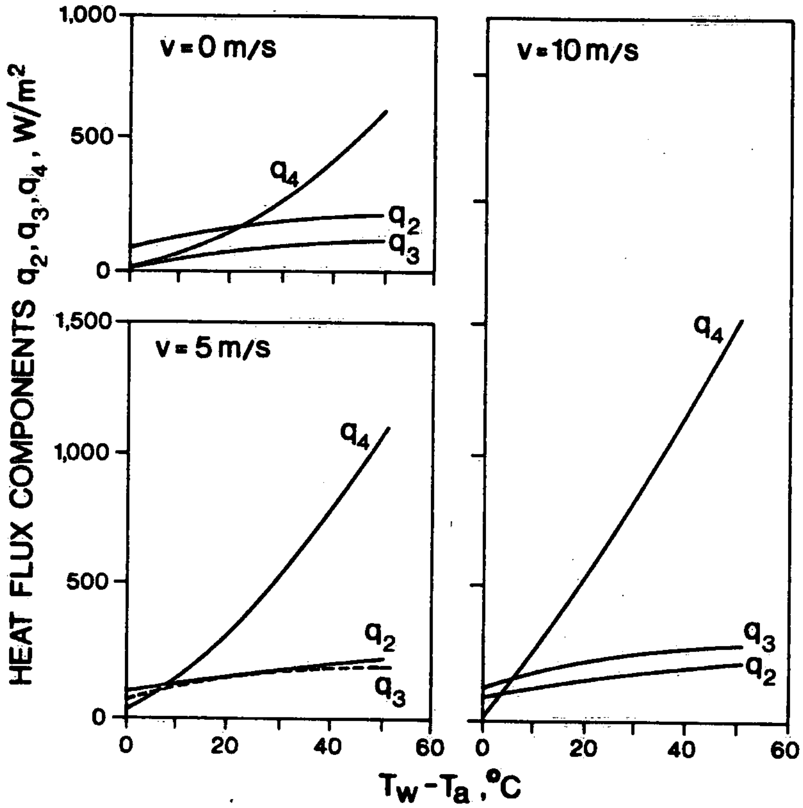


Fig. 1. Major components of heat loss rate during freeze up, as calculated by Dingman et al (1968) for clear sky and 50% relative humidity.

q_2 = long wave radiation loss

q_3 = evaporative loss

q_4 = loss by convection and conduction (usually much smaller than convective loss)

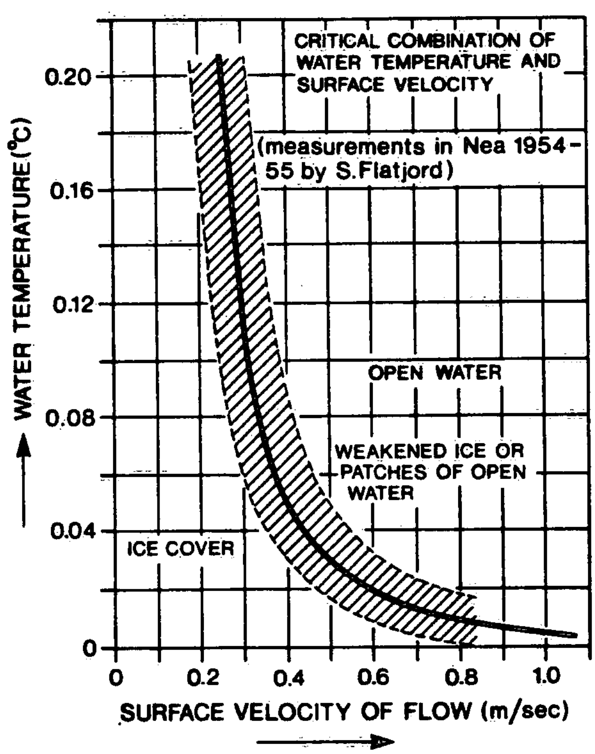


Fig. 2. Conditions for the formation of border ice during period of intense cold. Measurements by S. Flatjord, as quoted in Prowse, 1987.

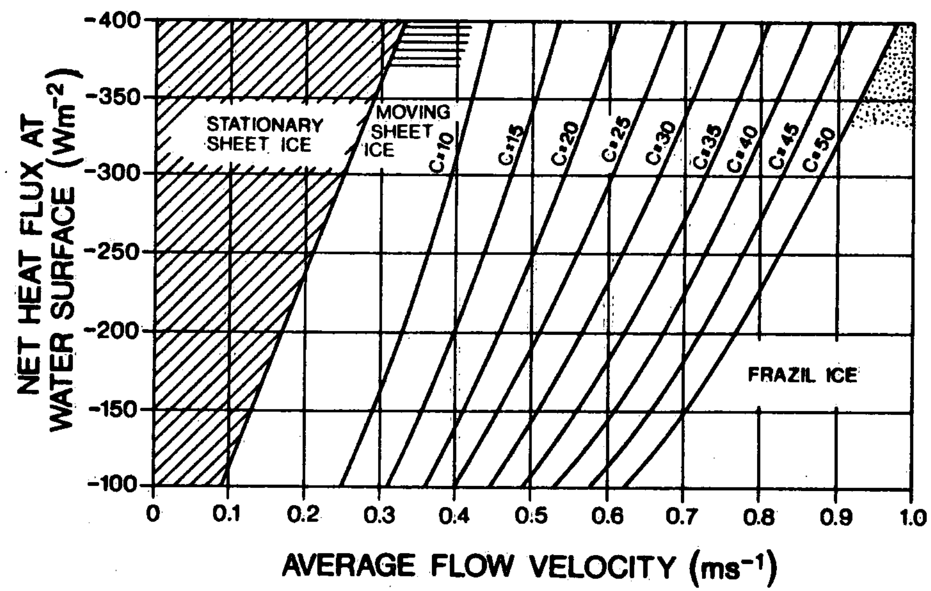


Fig. 3. Conditions for formation of various ice types (from Matousek, 1984; for wind speed at two metres above surface = 0.5 ms^{-1}; average water temperature = 0°C; surface width in wind direction = 130 m; C = chezy coefficient = $R^{1/6}/n$, n = Manning coefficient).

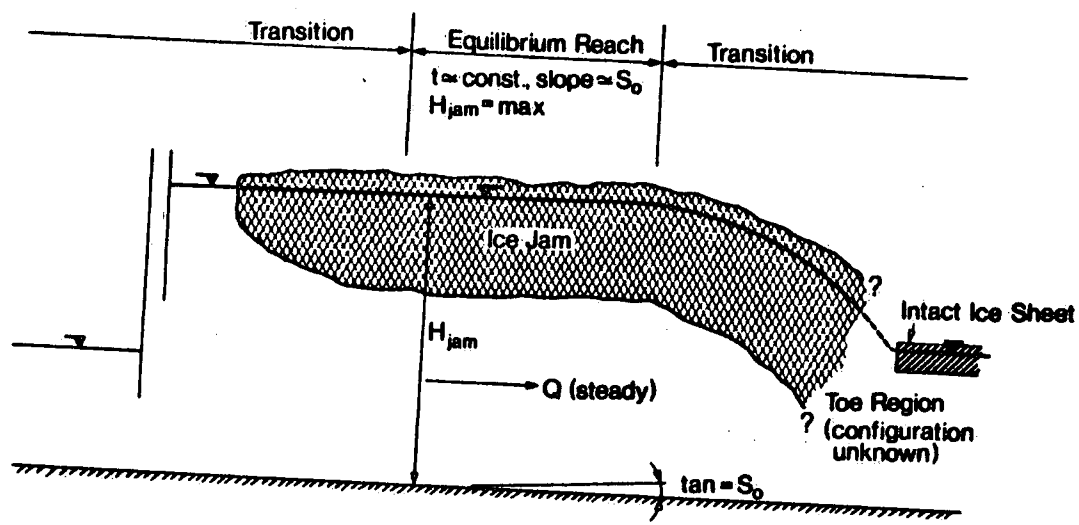


Fig. 4. Profile of a floating jam with an equilibrium reach.

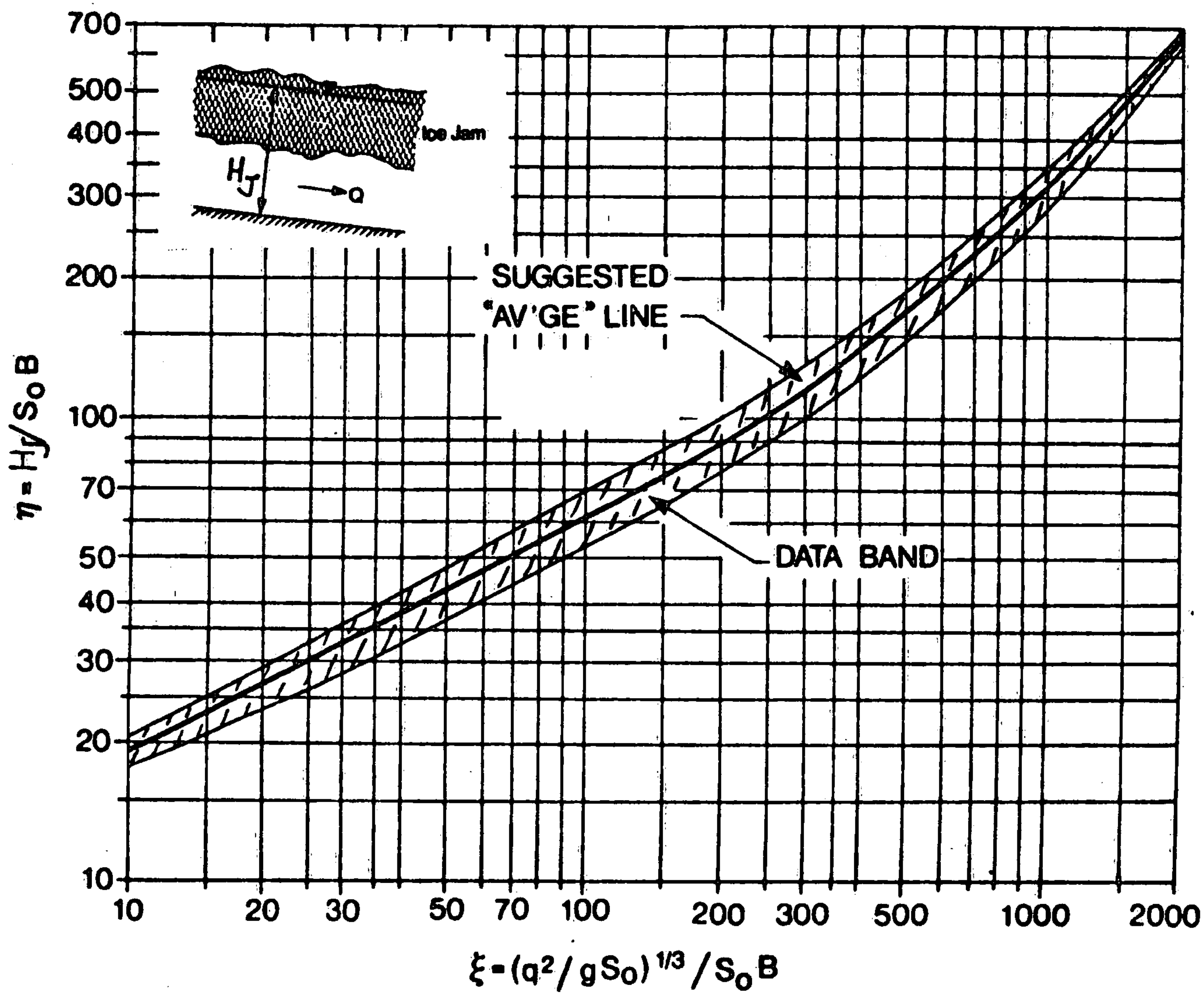
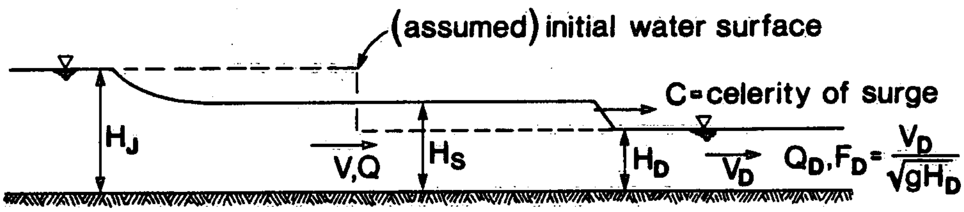
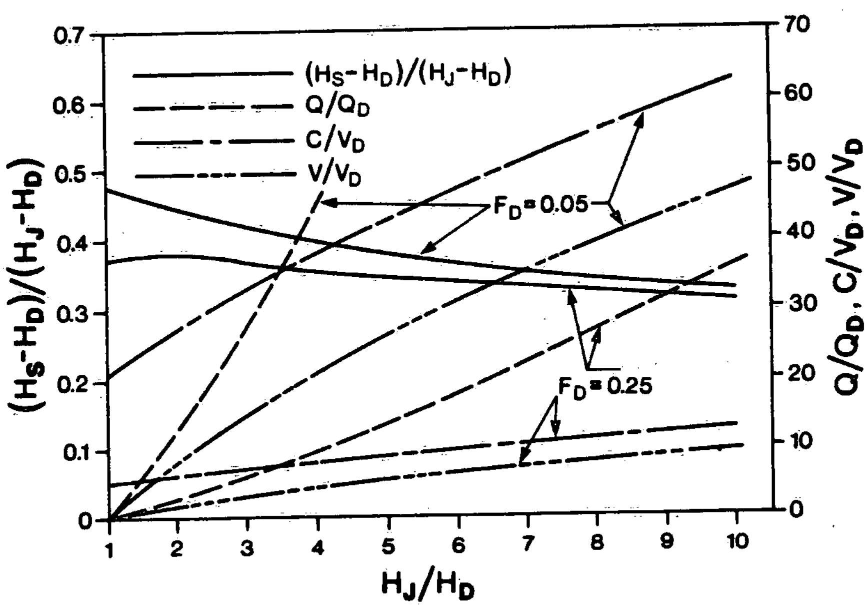


Fig. 5. Dimensionless relationship to compute breakup jam equilibrium levels (from Beltaos 1986; Q = discharge; B = channel width; q = Q/B; S_o = channel slope).



(a) Definition sketch. Note idealized shape of initial water surface, assumed for simplicity.

H_D, V_D, Q_D, F_D = respectively flow depth, velocity, discharge and Fronde number downstream of the jam prior to release.



(b) Illustrative theoretical results. Note usual range of H_J/H_D is 1 to 3.

Fig. 6. Surge characteristics, as determined by theory of Henderson and Gerard (1981). The effects of channel slope and bed friction are neglected.