CHAPTER 13

Ice Effects on Sediment Transport in Rivers

Robert Ettema

13.1 INTRODUCTION

The winter cycle of river-ice formation seasonally grips many rivers in large areas of the Northern hemisphere. The cycle affects river-channel capacity to convey water and sediment, it may aggravate riverbank erosion, and it may perturb the stability of alluvial channels. The severity with which the cycle of river-ice formation affects sediment-transport dynamics for rivers depends on a combination of factors related to the cycle's duration and coldness (usually expressed as accumulated degree-days of freezing). Figure 13-1 indicates the extent of the Northern hemisphere that annually experiences at least 1 month of average air temperature below 0 C.

Of major importance is the seasonal availability of water flow. Under natural conditions in many rivers, the winter cycle of ice is accompanied by a decline in water runoff and channel flow. Rates of sediment supply and channel transport diminish commensurately. Runoff and channel flow subsequently increase during spring thaws, and it is then that ice-cover effects on sediment transport become significant. For many flow-regulated rivers subject to the ice cycle, though, ice effects on sediment transport and alluvial-channel behavior are of special interest. Substantial flows may occur while such rivers are ice-covered in winter.

Also important are the materials composing the bed and bank of a river. Diverse other factors, such as north-south river flow orientation and snowfall, also can exert significant influences. The overall impacts of all these factors on sediment transport and channel morphology vary widely from one river to the next and differ from reach to reach along a river. The impacts may be distinct and clearly observable, for rivers in permafrost or annually subjected to severe ice runs following ice-cover breakup in spring. They may be obvious from stunted riparian vegetation, scarred trees, or gouged channel features. They may be subtle and blurred by the inherent complexities and apparent irregularities of alluvial-channel flow. They also may be intermittent, being significant at one site on one occasion, but not the next. A good deal of the variability in ice impacts is attributable directly to variability in flow conditions.

Ice effects on sediment transport may be noticeable over varying scales of time and channel length. On the scales of months and of miles of channel, for instance, ice alters the relationship between flow rate, flow depth, and sediment transport rates. As it forms, an ice cover usually increases and redistributes a channel's resistance to flow and reduces its overall capacity to move water and sediment. In a sense, because the channel's bed roughness does not actually increase (in fact it may decrease; Smith and Ettema 1997), the effect of ice-cover presence on channel morphology may be likened to the effect produced by a reduction in energy gradient associated with flow along the channel. More precisely, it may be likened to a change in thalweg geometry; the additional flow energy consumed in overcoming the resistance created by the cover offsets a portion of the flow's energy that the channel dissipates by thalweg lengthening or bifurcation. This sort of postulation, though fun and possibly sound theoretically, may be difficult to verify practically, because ice covers vary in length, thickness, and roughness along most rivers. The fact remains that, at present, scant data exist for rivers.

On the local scale, an ice cover over a short reach may redistribute flow laterally across the reach, accentuating erosion in one place and deposition in another place. Such local changes of the bed may develop during the entire cycle of ice formation, presence, and release. They may develop briefly, lasting slightly longer than the ice cover, and disappear shortly after the cover breaks up. Or they may trigger a change that persists for some time. In any event, they should be verifiable from a site investigation.

Ice may dampen or amplify erosion processes locally. Obvious damping effects of ice are reduced water runoff from a watershed, cementing of bank material by frozen

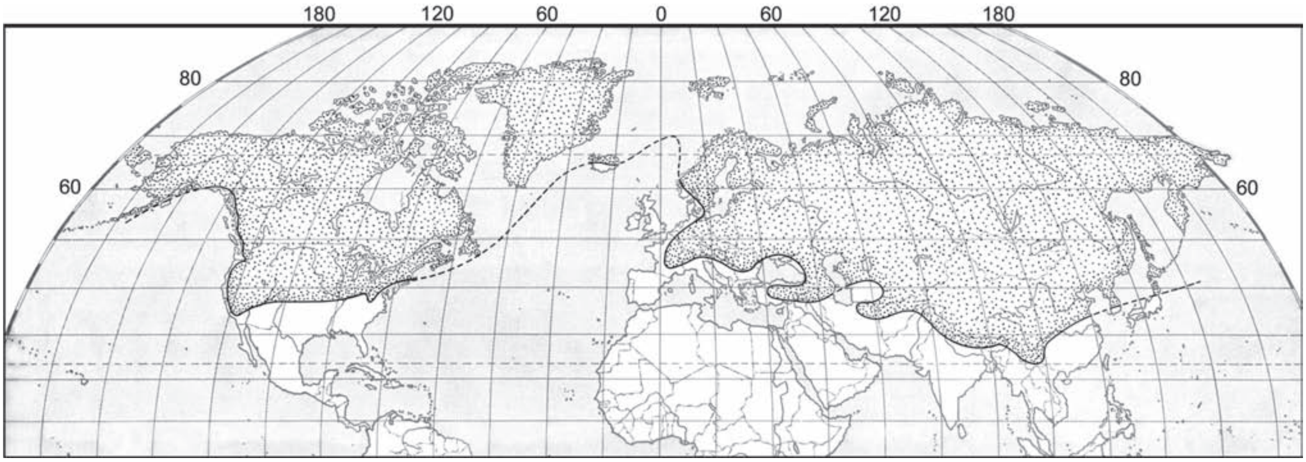


Fig. 13-1. Area of Northern Hemisphere that experiences at least one month per year with average air temperatures less than 0 C.

water, and ice armoring of bars and shorelines by ice-cover setdown with reduction in flow rates. Yet ice may amplify erosion and sediment-transport rates, notably during the surge of water and ice consequent to the collapse of a large ice jam.

In recent years, growing interest in the wintertime management of reservoir-regulated rivers that become ice-covered, especially the winter environments of such rivers, has made it necessary to better understand and model flow and transport processes in ice-covered alluvial channels. The need has become especially acute for river reaches in which flow regulation results in significantly larger wintertime flows than occurred during preregulation conditions. For such situations it is necessary to develop more accurate estimates of flow stage, quantity of sediment conveyed, and possible changes in channel morphology. Even for essentially unregulated or wild rivers, such as the Yellowstone River shown in Fig. 13-2, it has become important to understand channel response to the winter cycle of ice.

The present chapter describes how the ice cycle may affect sediment transport, locally as well as over long reaches. It is necessary to point out that the literature dealing with ice effects on sediment transport and channel morphology is not extensive. Moreover, what exists contains a fair amount of hypothesis and conjecture. Inevitably, therefore, this chapter also contains its share of hypothesis and conjecture. An unavoidable difficulty is that ice can have various and, at times, apparently contradictory effects. General conclusions about the net effects of ice are not at all straightforward to state, except to say that ice effects are closely related to velocity and elevation of flow; i.e., higher flows incur higher impacts under ice-covered conditions than under open-water conditions.

This chapter does not address the influence of permafrost on sediment transport. Permafrost is an important factor affecting riverbank and channel stability of high-latitude rivers. Scott (1978) and Lawson (1983), for example, provide some insights into channel behavior in permafrost. Johnston (1981) and Andersland and Anderson (1990) usefully describe the geotechnical properties of permafrost.

This chapter begins with an introductory description of the typical cycle of ice formation, ice effects on flow distribution, and ice-cover breakup in rivers (names commonly used for the various ice formations are introduced in italics). It then briefly describes how ice can directly entrain and transport sediment from the beds of certain rivers. Subsequently, it goes on to discuss the typical mechanisms whereby ice and cold water influence sediment transport by flow in rivers. The latter portion of the chapter addresses river-ice influences on

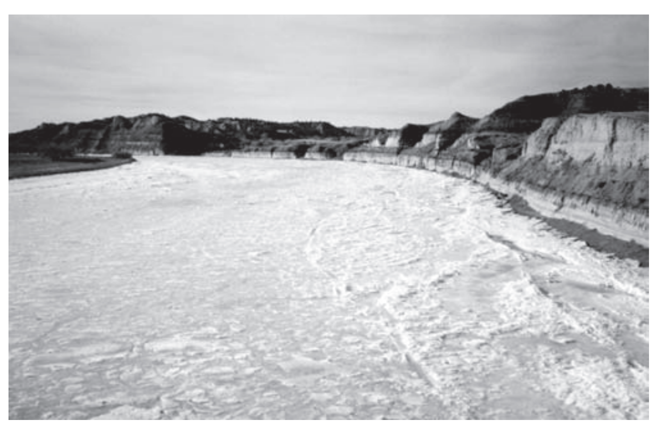


Fig. 13-2. The Yellowstone River, Montana, under an ice cover, whose formation, presence, and eventual breakup significantly influences sediment-transport dynamics, channel-thalweg location, and riverbank erosion.

channel stability. Of particular interest, in this regard, are rivers whose inflow is regulated by upstream dams.

13.2 ICE FORMATION

During autumn and into winter, river water cools. In cold regions, such as indicated in Fig. 13-1, it usually cools to the water-freezing temperature, or momentarily to a fraction of a degree below it (supercooling is needed to nucleate or initiate ice growth), whereupon ice rapidly forms. For a river whose inflow is regulated by a large reservoir, water temperature decreases with downstream distance of flow, and initial ice-cover formation develops commensurably at some distance downstream of the reservoir.

Ice-cover formation over a river comprises several main processes, which Fig. 13-3 (from Matousek 1984) usefully summarizes in terms of bulk velocity of flow, U, and heat flux to air, ϕ. The ice terminology in Fig. 13-3 is explained further in the ensuing sections of this chapter. Implied in this figure are the influences of vertical and lateral mixing within the flow, as well as the strength of thin newly formed ice. As water cools below 4 C, it becomes lighter, and thereby more difficult to mix within the body of flow.

One formation process is static or thermal, and could be called the *bankfast-ice* process. It occurs for flows of negligibly small surface velocity and is evident as ice growth outward from riverbanks. *Frazil-ice* formation starts with the supercooling of water over some portion of flow depth. *Skim ice* may form on flows whose surface velocity and turbulence levels are sufficiently low so that only a surface layer of water supercools, resulting in thin sheets of ice. As flow velocity and turbulence levels increase, the initial formation of ice becomes more dynamic. The full depth of water at some reach may supercool, so that frazil-ice crystals form throughout the flow. Frazil-ice formation usually dominates ice formation on alluvial rivers whose flow has sufficient velocity to move bed sediment. A given reach of river may undergo all three forms of ice growth, depending on the distribution of flow velocity upstream of and through the reach.

For river flow in a watershed unregulated by dams, factors related to channel size and air-temperature variation with altitude and latitude determine rate of water cooling and where ice first appears and gradually envelops a channel. Though exceptions exist, ice first forms in the upper reaches of a watershed for most rivers that drain toward the south (e.g., the Mississippi River). For northward draining rivers (e.g., the Red River of the North, the Mackenzie River), or rivers with more or less east-west orientations (e.g., the Yellowstone River, the Yukon River), the sequence of ice formation occurs in a more complicated manner along the length of the river.

An important factor influencing first ice formation in a channel whose flow is regulated by an upstream reservoir is the temperature of the water released into the channel. In most situations, the reservoir changes the temperature of the flow entering the channel; besides storing water volume, a reservoir stores heat. During freeze-up conditions in late fall and winter, the flow entering the channel is warmer than the flow in the channel prior to construction of the reservoir. Consequently, the reservoir likely will cause ice formation to

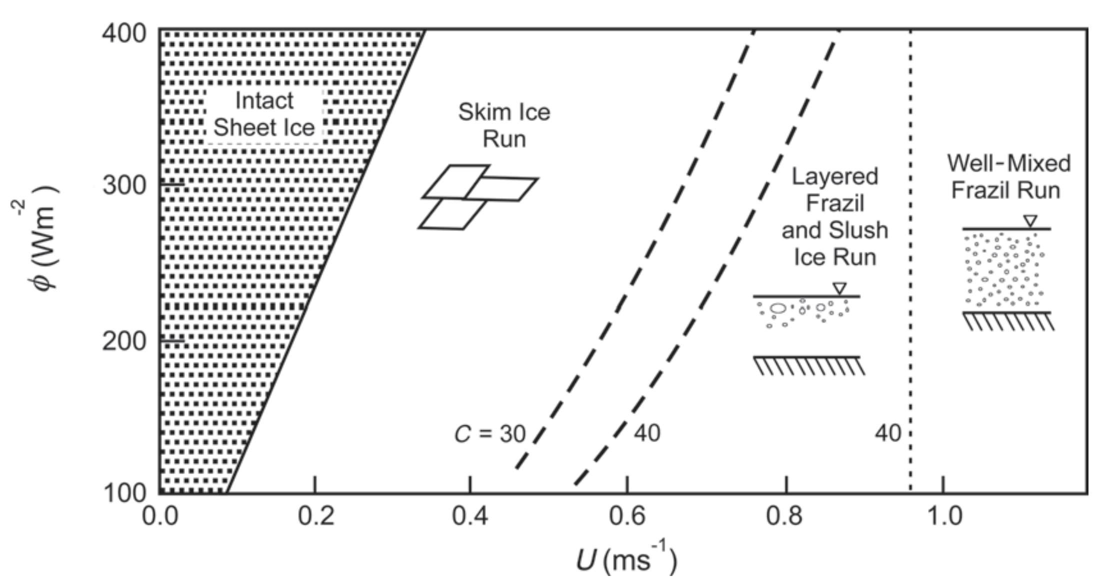


Fig. 13-3. Types of initial river-ice formation as a function of flow velocity and surface heat-loss rate; C is Chezy coefficient, and ϕ is rate of heat loss per unit area of river water surface (Matousek 1984). Larger flow velocity, U, results in greater mixing and cooling of flow over its full depth.

begin further downstream along the channel than it did prior to construction of the reservoir. The thermal influence of a reservoir on ice-cover formation can be demonstrated quite readily. If, prior to construction of the reservoir, the flow entering the channel reach was at the freezing temperature of water (0 C), ice potentially could begin forming throughout the full length of the channel. For example, if a 3-m-deep flow of 4 C water were released with an average velocity of 1 m/s and exposed to −20 C air under representative conditions of heat loss (say, 20 W/m^2 / C), the flow would travel almost 117 km downstream from the reservoir before cooling to 0 C. If the initial temperature of the water leaving the reservoir were 1 C, the distance would be reduced to about 30 km. Therefore, as the water in the reservoir's water cools during winter, the ice cover on the river may progress further upstream. Water density is greatest at 4 C, and therefore water at the elevation of a reservoir's outflow conduit, usually placed low through the dam, is likely to be at this temperature.

Cold, clear, windy nights are especially conducive to ice formation. During such nights, rivers lose heat at maximum rates to the atmosphere by means of long-wave radiation, convection, and evaporation. Consequently, it is common for ice to form, or at least to form at its greatest rate, during the night.

13.2.1 Bankfast Ice

As can be seen from the river view shown in Fig. 13-4, bankfast ice (also called border ice) usually is the first type of ice to appear along a river. It forms in low-velocity zones along banks. The top layer of the water adjacent to the bank mixes minimally with lower layers and soon becomes supercooled in frigid air, while water elsewhere is still above the freezing temperature. Ice fragments in the air and at the riverbank nucleate the supercooled water at the surface. The nucleated water propagates an ice sheet on the water surface outward from the bank. The edge of the ice sheet eventually extends to a zone of turbulent water, whereupon its further progress depends on thermal atmospheric exchange. The growth does not stop just because the water is above the freezing temperature, though it slows. It continues growing by virtue of net heat loss of water fringing the bankfast-ice edge. Bankfast-ice extension accelerates when drifting frazil slush and small pans lodge against it. The slush and pans fuse in rows to the dendrite crystals extending from the bankfast-ice edge, and they may form successive layers in the outward progressing border ice.

Bankfast-ice growth is a prominent ice-formation process in small rivers and streams with mild slopes. Together with skim ice, it is the static type of ice growth that occurs in lakes during calm but frigid weather. In the context of bank-erosion concerns, the effects of border ice on bank-material strength and loading are not well understood. For instance, not much is known about how bankfast-ice growth affects freezing of groundwater within a riverbank.

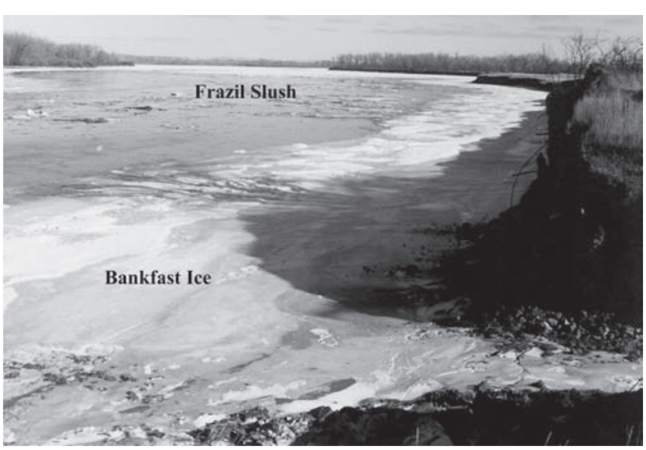


Fig. 13-4. Bankfast ice formed along a bank of the Missouri River. Frazil-ice slush and pans drift in center channel.

13.2.2 Skim Ice

When surface velocities are low, the surface layer of flow may become supercooled and spawn frazil ice, which rises and forms fragile, thin sheets of skim ice (e.g., Matousek 1984; Ashton 1986). Marcotte (1984) reports large sheets of skim ice forming when surface velocities of flow along the St Lawrence River were about 0.3 m/s; he reports that, in very cold weather, skim ice may form at surface flows with velocities of about 1.0 m/s. Sheets of skim ice drift until they gently lodge against each other along a river. The river then quickly freezes over completely. Skim ice and bankfast ice are common ice forms on rivers and streams whose slopes are sufficiently mild so that flow velocities are of the magnitude ranges tentatively indicated above.

13.2.3 Frazil Ice

For fully turbulent flow, frazil-ice formation begins with the formation of frazil-ice crystals throughout the depth of flow in an ice-generation zone. It is an especially striking and dominant feature of river behavior in cold regions.

Frazil ice appears quickly in a flow that supercools to a fraction of a degree below the freezing temperature of water, i.e., nominally about −0.01 to −0.1 C. As frazil crystals form, the latent heat of fusion they release gradually raises the water temperature to 0 C. During this period, the frazil is in what is termed the "active" state, in which it fuses readily with solid objects that it contacts (e.g., other frazil ice crystals, sediment on the riverbed, boulders, and some aquatic plants). For a flow-regulated river, the zone of active frazil formation may be fixed and extend only a few hundred feet, producing frazil conveyed downstream by the flow. The continuous variation in weather conditions in nature (notably, fluctuations in air temperature, wind speed, and net heat loss by means of radiation) causes the zone to shift. Lowering air temperature or water flow rate, for example, causes the zone to move upstream. Supercooling could occur at the same river site for several days, depending on daily fluctuations in weather and flow. As an ice cover forms and progresses downstream along an unregulated river, the zone of frazil-ice production may also move downstream.

Frazil crystals grow rapidly in size, fuse to each other, agglomerate, and (owing to ice buoyancy) rise to the water surface if able to drift for a sufficient distance of flow. When initially in supercooled water, frazil crystals fuse to almost any solid boundary in the flow. For instance, they may fuse to the river bottom, forming an accumulation termed *anchor ice*. As frazil drifts, it rises to the water surface, agglomerates, crusts over, and forms ice pans, which have a hard, flat circular top and an approximately hemispherical accumulation of slush below. At this stage, the water no longer is supercooled and the frazil is termed inactive frazil; it has lost its propensity to fuse readily.

Long reaches of rivers may become covered with drifting *slush*, *pans*, and *floes* formed of fused pans. Figure 13-5 (adapted from Michel 1971) illustrates the genesis of an ice cover formed primarily from frazil ice. In deep sections with relatively low surface velocity, or in other locations with low surface velocities, the ice coverage concentrates. The pans and floes drift with the flow until they become congested (such as in a traffic jam) or lodge against some constriction. Once cover has started, it progresses upstream rapidly as a juxtaposed layer of pans and floes cemented with frazil slush. It is typical for ice covers on large rivers to progress upstream at a rate of about 40 km per day in this manner (e.g., Michel 1971). Alternately, a pile-up of ice may occur and form what is termed a freeze-up jam. *Freeze-up jams* retard flow and raise water levels, possibly causing flooding upstream of the jam toe. Several flow-related variables influence the upstream progression of a level cover comprising juxtaposed pans and floes. However, an approximate rule of thumb (e.g., Michel 1971; 1978; Ashton 1986) is that a level cover may develop when the Froude number for flow at the site is about 0.1 or less; i.e., the Froude number $= U/(gY)^{0.5} < 0.1$, in which U = bulk flow velocity, Y = flow depth, and g = gravitational acceleration. For typical rivers, it is easier to use simple velocity criteria; e.g., frazil slush passes under the front of an ice cover when flow velocity exceeds about 0.6 m/s, and frazil pans will go under when velocity exceed about 2 m/s. The cover still may progress upstream when ice passes under its front if the rate of ice arrival at the cover front exceeds the rate at which ice is subducted beneath the front.

When the upstream front of the cover reaches a high-velocity section of a river, large amounts of slush and pans are forced under the front and conveyed beneath the cover. The slush sometimes forms clusters and *granules* or *pebbles* conveyed long distances under ice covers, being transported as a form of "bed load" of frazil that rumbles along the cover underside (Shen and Wang 1995). The granules, as well as slush and small pans, may come to rest and accumulate in zones of lower velocity beneath the cover. Chacho et al. (1986) describe similar transport of frazil along the underside of the ice cover of the Tanana River, Alaska.

Large accumulations of ice may develop under the cover and be resistant to shoving. In some situations, they may form a feature known as a *hanging dam*. Ice moving under

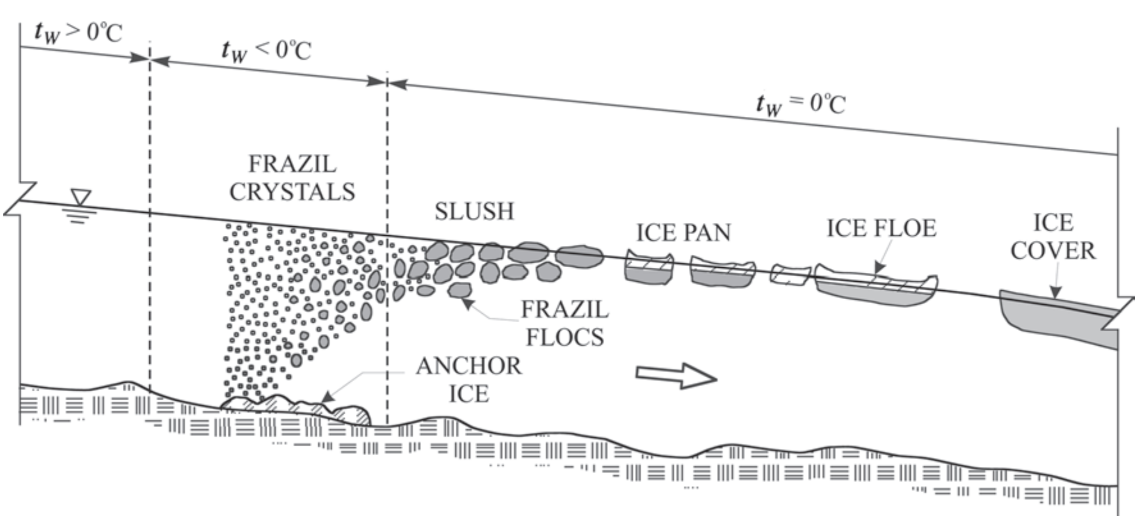


Fig. 13-5. The genesis of frazil ice in a river or stream. The water cools until slightly supercooled, whereupon frazil ice crystals rapidly appear, agglomerate as slush, develop as ice pans, which then may align juxtaposed as an ice cover. t_w = temperature of water.

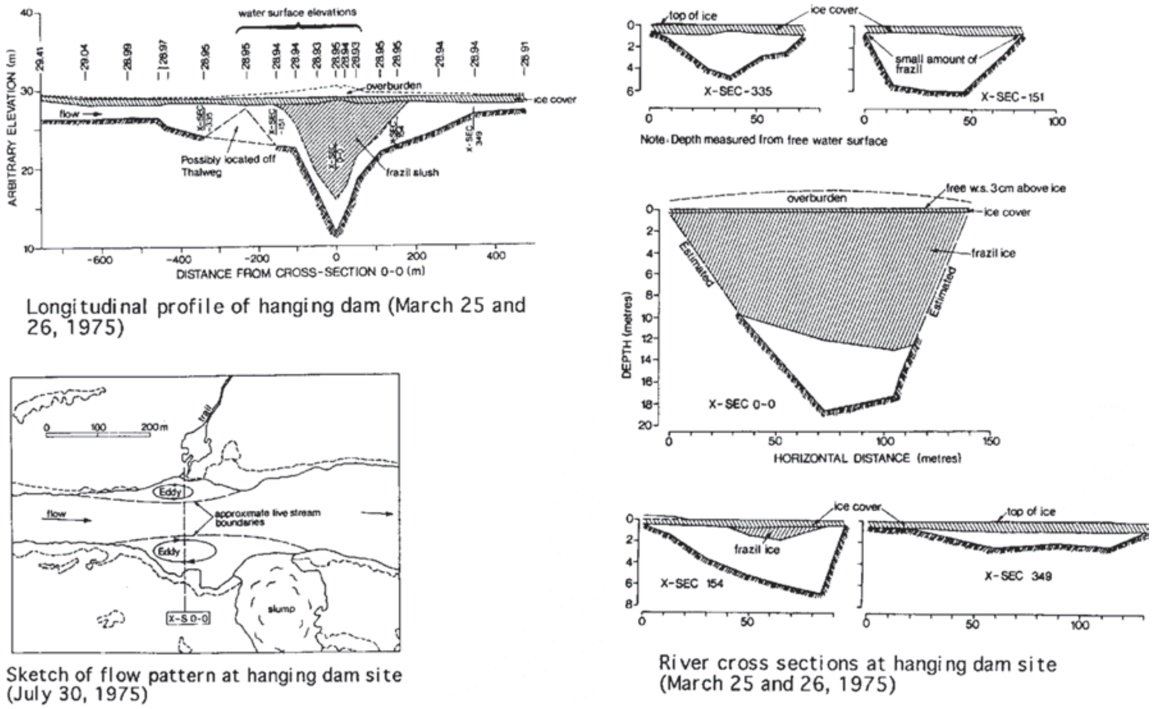


Fig. 13-6. A hanging jam of frazil ice may develop under an ice cover when flow velocities exceed those needed to form an ice cover of juxtaposed pans of ice. This example, taken from Beltaos and Dean (1981), shows a hanging dam in the Smoky River, Alberta.

the cover progressively accumulates in locations of reduced flow velocity, concentrating the flow velocity so that it locally scours the river's bed (a later section of the chapter further discusses this concern) and increases flow area. The hummocking of an ice cover can be a clue to the presence of a hanging dam. As hanging dams and similar accumulations increase in size, they increase flow resistance, raise water level, reduce and possibly redistribute flow velocity, and enable the cover to continue progressing upstream. Figure 13-6, taken from Beltaos and Dean (1981), depicts typical aspects of a hanging dam in the Smoky River, Alberta.

For steep, highly turbulent streams, another form of dam building occurs. Weirs of anchor ice (frazil ice bonded to the bed, not the ice cover) may extend up from the streambed, reducing flow velocity and enabling the cover to progress upstream. The anchor-ice weirs retard the flow and eventually help a cover form over the flow.

In relatively steep, swift-flowing channels, frazil ice may not develop to the level cover of juxtaposed pans or covers with hanging dams. Instead, the higher flow velocities associated with steeper channels, pans, and slush, sometimes mixed with snow, form a jumbled accumulation known as a freeze-up jam. Such jams may be free-floating or partially grounded on the bed. The remnant of such a jam in a gravel-bed reach of the Yellowstone River, Montana, is depicted in Fig. 13-7. The jam clogged much of the reach, especially in shallower, slower current portions to the side of the river's thalweg.

A cover of pans and slush solidifies contiguously between the ice pieces and may thicken thermally. The contiguous solidified cover resists the hydrodynamic drag exerted by the water and the streamwise component of the cover's weight. The cover may locally buckle, shove, hummock, and bummock at weak spots as the cover progresses upstream, the flow rate fluctuates, and/or air temperature changes.

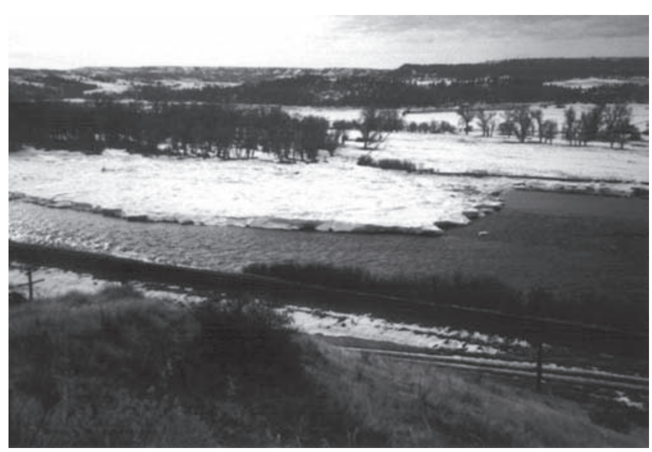


Fig. 13-7. The remnants of a freeze-up jam in the Yellowstone River, Montana. The jam comprised frazil slush and pans mixed with snow and is partially grounded.

13.3 ICE-COVER EFFECTS ON FLOW DISTRIBUTION

An ice cover imposes an additional resistant boundary that decreases a channel's flow capacity and vertically redistributes streamwise velocity of flow in a channel. If the cover is free-floating, it may reduce the erosive force of flow in the channel and thereby reduce rates of sediment transport. However, cover presence also may laterally redistribute flow, usually concentrating it along a thalweg. If the thalweg lies close to one side of a channel, flow concentration may locally increase bank erosion and channel shifting. On the other hand, if the thalweg is more or less centrally located in a channel, the cover may reduce bank erosion and channel shifting. Additionally, if the full cover is fixed to the riverbank, it may increase locally flow velocities and rates of sediment transport.

The variability of flow response to ice cover makes it difficult to draw simple overall conclusions about ice-cover effects on a river's bed and banks. The net effects will vary from site to site.

If the flow rate and channel slope are assumed constant, the main individual effects of a uniformly thick ice cover on a straight uniformly deep alluvial channel are as follows:

1. Raised water level (ice-covered depth exceeds open-water depth for the same flow rate);
2. Reduced bulk velocity of flow (discharge/flow area);
3. Reduced drag on the channel bed;
4. Reduced velocity of secondary currents (i.e., currents associated with transverse circulation of flow in the channel);
5. Reduced rates of bed-sediment transport; and
6. Altered size and shape of bed forms (notably dunes).

The effects are evident in the comparison of Figs. 13-8(a and b) and the ensuing explanation, which considers cover effects on flow distribution in fixed-bed channels. Section 13.6 discusses flow and sediment transport in ice-covered alluvial channels.

13.3.1 Ice-Cover Influence on Vertical Distribution of Flow (Fixed Bed)

The direct effect of imposing a level ice cover on a two-dimensional open-water flow is to increase the wetted perimeter of the flow substantially. For a wide channel, the wetted perimeter is almost twice that for the open channel flow. The usual consequence is increased water depth for constant discharge and bed slope, as indicated in Figs. 13-8(a and b). Because roughness characteristics of the ice-cover and the bed likely differ, the influence of the roughness of the bed and the ice-cover underside on velocity distribution and flow resistance must be taken into account. However, doing so accurately is not straightforward.

For the same uniform two-dimensional flows (one open-water, the other covered, as in Figs. 13-8(a and b)) having

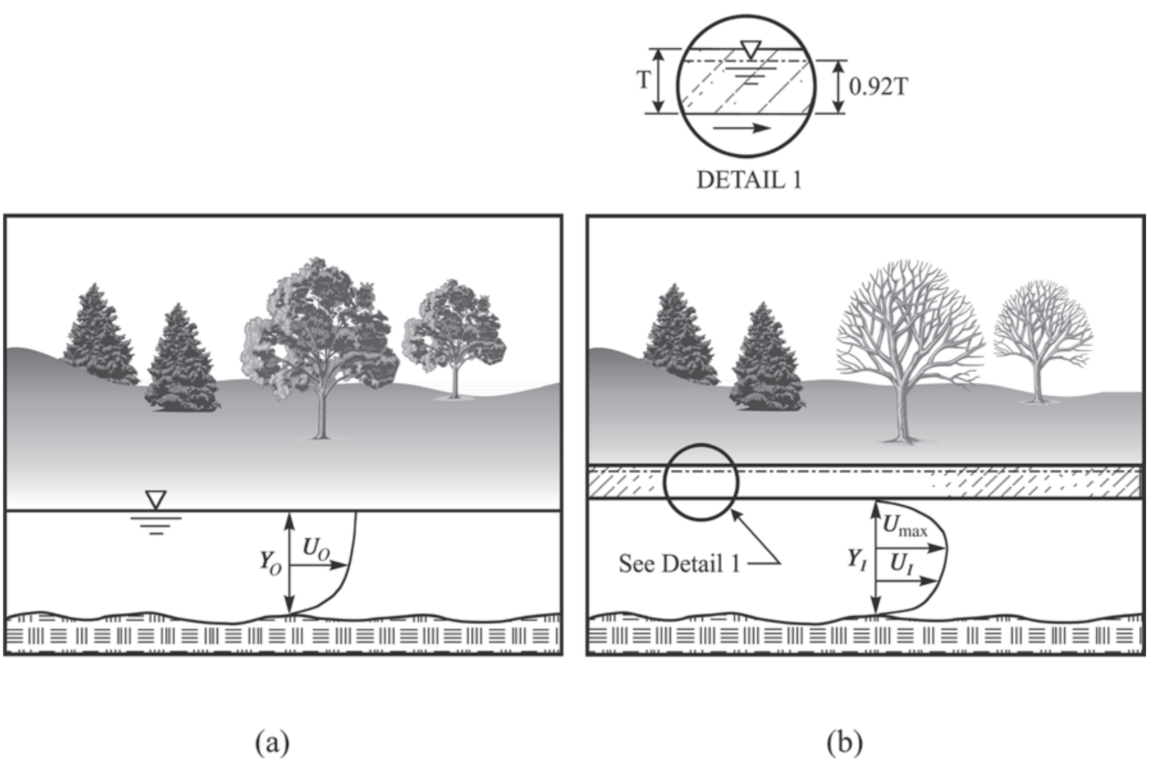


Fig. 13-8. Presence of a free-floating ice cover usually increases flow depth and redistributes flow.

the same unit discharge and energy slope, the ratio of flow depths is

$$\frac{Y_I}{Y_O} = \frac{U_O}{U_I} = \left(2\frac{f_I}{f_O}\right)^{1/3} \tag{13-1}$$

in which Y and U define the flow depth and the bulk velocity, respectively; f is the Darcy-Weisbach resistance factor; and subscripts O and I refer to open-water and ice-covered flows, respectively. Because the overall resistance coefficient for the ice-covered flow, f_I, exceeds 0.5-f_O (where f_O is the resistance coefficient for the same discharge during open-water flow), $Y_I > Y_O$. Typically, flow depth increases by about 10 to 30%; i.e., the covered flow is about 10 to 30% deeper than the open-water flow for the same discharge. Bulk velocity of flow decreases by the same amount. The actual piezometric water level in the channel would be about 0.92 times the ice-cover thickness, T, above the cover underside; the density ratio of solid ice and water is 0.92. The cover floats with a freeboard of about 0.08T above the water level. The freeboard can be suppressed or even slightly negative if there is a thick snow layer on top of the ice cover.

The resistance factor for covered flow, f_I, is a composite value expressing the total resistance exerted by the bed and the cover; i.e., for two-dimensional flow,

$$\mathrm{f_I} = \frac{4\tau_{\mathrm{tot}}}{\rho \mathrm{U_I^2}} = \frac{4\left(\tau_b + \tau_i\right)}{\rho \mathrm{U_I^2}} \tag{13-2}$$

in which

τ_{tot} = combined flow resistance exerted by the bed and the ice cover;
ρ = water density;

and subscripts b and i refer to the channel bed and ice-cover underside, respectively. Equation (13-2) implies that a change in resistance at one surface will alter U_I and thereby alter the flow resistance exerted by the opposite surface.

The customary practice (e.g., Michel 1978; Ashton 1986; Beltaos 1995) is to calculate flow resistance in ice-covered channels using what is termed the two-layer hypothesis, whereby flow resistance is taken to be a linear composite of flow resistance attributed to flow drag along the bed and ice-cover underside. In accordance with this hypothesis, each resistance coefficient or drag contribution is assessed independent of the other. The resistance coefficients are related to boundary roughness (bed or ice underside) normalized with the part of the total flow depth extending from the pertinent boundary to the elevation of the velocity maximum. This approach entails use of the Sabaneev equation, a semiempirical approximation proposed by A. A. Sabaneev (Nezhikovskiy 1964),

$$n_I = n_b\left[\frac{1 + \left(n_i/n_b\right)^{3/2}}{2}\right]^{2/3} \tag{13-3}$$

in which

n_b and n_i = values of Manning's resistance coefficient associated with the bed and the ice cover, respectively; and
$n = R^{1/6} f^{0.5}/(8g)^{0.5}$, with
R = hydraulic radius.

The two-layer hypothesis is inadequate for estimating bedform geometry, flow resistance, rates of sediment transport, and dispersion processes. Significant physical inaccuracies arise in partitioning covered flow in accordance with the two-layer hypothesis and applying Manning's equation to estimate flow resistance for each part. Flow resistances, both at the bed and at the ice cover, directly alter distribution of flow velocity. They affect the length scales and intensities of turbulence across the full depth of flow. Flow resistance is not simply the sum of flow drag determined from linear variation for each boundary, independent of the other boundary. In effect, there occurs an interactive "cross-torque" between the ice cover and the bed.

Even when the two roughnesses are identical, strictly speaking it is not physically meaningful to partition the flow at the plane of maximum velocity. Though, in this case, the plane may coincide with the plane of zero shear stress and mid-depth, the upper limit of the turbulence structures in the flow scales with the full flow depth, and turbulence diffuses and interacts across the flow. Differences in ice-cover and bed roughnesses offset the elevations of maximum velocity and zero shear stress. The offset increases when the roughness of one boundary is markedly greater than that of the other boundary.

The wind tunnel experiments carried out by Hanjalic and Launder (1971) and Reynolds (1974), for flow through a duct with top and bottom boundaries of differing roughness, and the flume experiments conducted by Gogus and Tatinclaux (1981) and Muste et al. (2000) are worth mentioning. These studies found, as expected, that the difference in top and bottom roughness shifted the position of the maximum streamwise velocity toward the roughest surface. In addition, the central region of the flow is characterized by strong diffusion of turbulent shear stress and kinetic energy from the rougher wall to the smoother one. The result is an appreciable offset of the plane of maximum velocity and zero-shear-stress plane. For a fully developed asymmetric flow, the noncoincidence of the surfaces of zero shear stress and mean velocity caused the production of turbulent kinetic energy to be negative over the central portion of the flow. In other words, a loss of turbulence energy occurs locally that affects velocity distribution over this portion. This loss of turbulence energy in the region where the smoother- and rougher-wall turbulence structures mix is attributable to the net interaction of Reynolds stresses of opposing sign.

Several recent studies report data on turbulence quantities for flow in ice-covered channels. One study (Muste et al. 2000), conducted with simulated free-floating ice cover and constant discharge in a laboratory flume, shows that cover presence decreases Reynolds stresses in the near-bed region and that rougher covers further decrease Reynolds stresses in the near-bed region. Turbulence was measured using laser-Doppler velocimetry. For a free-floating cover, such reductions in Reynolds stresses follow from overall reductions in bulk velocity of flow. Cover presence, however, did increase Reynolds stresses near the top of the flow. Martin and Roy (1999) report field measurements of turbulence for the Sainte-Anne River, Quebec. Their measurements, taken using an electromagnetic velocity meter, indicate that a damping of larger-scale turbulence structures occurs. Sukhodolov et al. (1999) used an acoustic-Doppler velocimeter to investigate turbulence structures in an ice-covered sand-bed reach of the River Spree, Germany. They found that the spatial scale of turbulence ejections emanating from bed forms was as large as flow depth, whereas sweeps scaled at about 0.7–0.8 flow depth.

13.3.2 Ice-Cover Influence on Lateral Distribution of Flow (Fixed Bed)

An ice cover imposes an additional flow-retarding boundary that decreases the flow-conveyance capacity of a channel and redistributes flow vertically and laterally. Vertical redistribution of flow is marked by flow depth increase (usually) and by null flow velocity at the cover underside. Lateral redistribution of flow, though, depends on how the ice cover forms, is attached to the channel banks, and thickens. It can be explained using the Darcy-Weisbach flow-resistance equation, written here for open-water flow in a channel of uniform depth,

$$Q_O = Y_O B\left(8gR_O S_O / f_O\right)^{1/2} = K_O S_O^{1/2} \tag{13-4}$$

in which

Q_O = flow rate;
B = flow width; and
K_O = unit conveyance.

Cover presence may laterally redistribute or concentrate flow in accordance with lateral variations in flow depth and/or ice-cover thickness. This impact can be illustrated in simple terms using an idealized channel comprising two bottom elevations of equal width, as in Figs. 13-9(a–d). Flow in such a channel may be described approximately in terms of two conveyance components, K_{O1} and K_{O2}, one component (1, 2) associated with each bottom elevation.

For constant flow, a free-floating, uniformly thick ice cover reduces the relative magnitudes of the two conveyance components. It smears flow over the full channel width, as $K_{I1}/K_{I2} < K_{O1}/K_{O2}$ (Fig. 13-9(b)). However, if the ice cover were fixed to the channel banks and thickened, the reverse

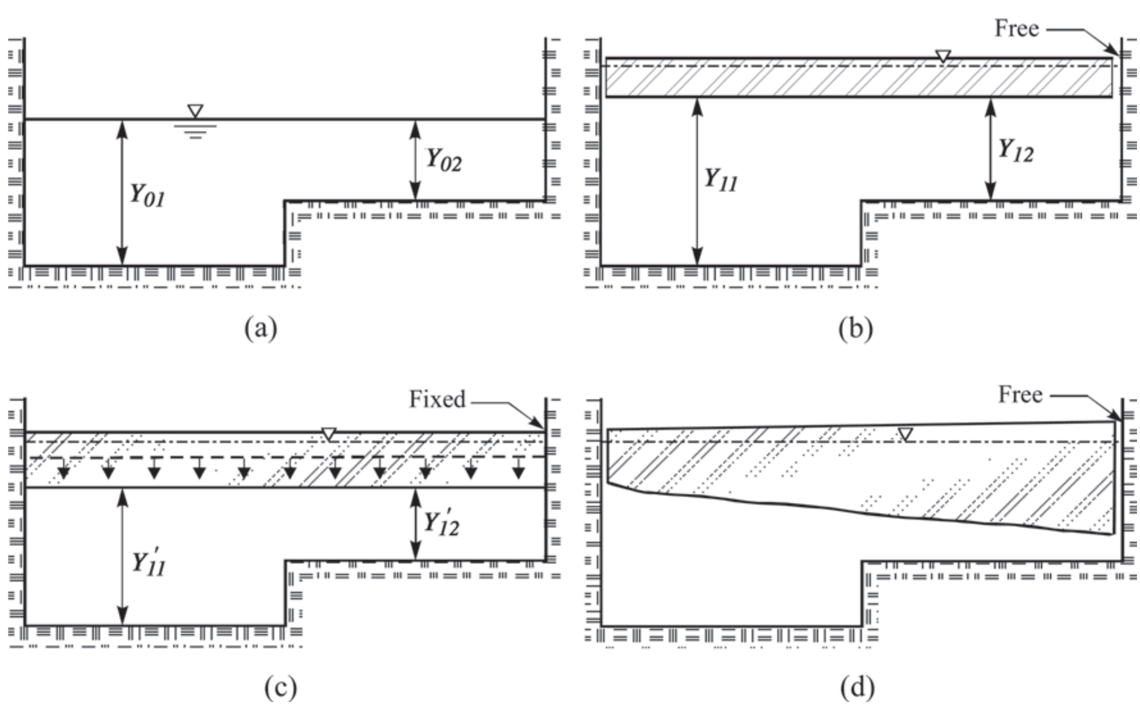


Fig. 13-9. An ice cover may reduce open-water proportions (a) of flow conveyance in lateral segments of a two-part compound channel if the cover is level and free floating (b); increase them if the cover is fixed and thickens (c); or, increase them if the cover is not uniformly thick (d).

would occur: $K_{I1}/K_{I2} > K_{O1}/K_{O2}$ (Fig. 13-9(c)), because flow depth decreases more in the shallower portion. Under this condition, cover squeezes or concentrates flow along a thalweg, where flow is deeper. If the thalweg lies close to one side of a channel (e.g., near the outer bank of a bend), such a concentration of flow may promote thalweg shifting and deepening. On the other hand, if the thalweg is located more or less centrally in a channel, a fixed cover may deepen or entrench the thalweg. An important further point is that the cover, by reducing flow through the shallow portion, may trigger further reductions in conveyance through the shallower portion by promoting ice accumulation (frazil slush or pans) and/or bed-sediment deposition there. Additional flow concentration is possible if the cover is not uniformly thick (Fig. 13-9(d)), if ice is grounded on the channel bed, or if shorefast/accumulated ice develops from one or both banks.

Lateral variations in cover thickness, however, may further concentrate flow in a channel of nonuniform depth and may override the more subtle effects described for a level ice cover. Significant lateral and streamwise variations in cover thickness may occur in channels with significant variations in flow depth and velocity. Because flow velocities decrease with decreasing flow depth, velocities usually are lower in regions of shallower flows and often in the wake of flow obstructions, such as bars. Ice covers whose formation involved substantial amounts of frazil-ice slush may become thicker in regions of shallower flow. Lower values of flow conveyance in those regions also result in relatively faster bankfast-ice formation. Also, because flow velocities are lower, ice (frazil slush and ice pieces) is less readily conveyed through those regions and is prone to accumulate. Figure 13-10 illustrates the accumulation of ice at a cross-section of the Tanana River, Alaska, at two times during winter (Lawson et al. 1986). That river is comparable to the lower Missouri River in flow rates, but is of steeper slope and more braided in channel morphology, and its flow is not regulated.

Further concentration of flow is possible if an ice cover is not free to float upward with increasing flow rate. Hydraulic analyses usually assume (e.g., Michel 1978; Ashton 1986;

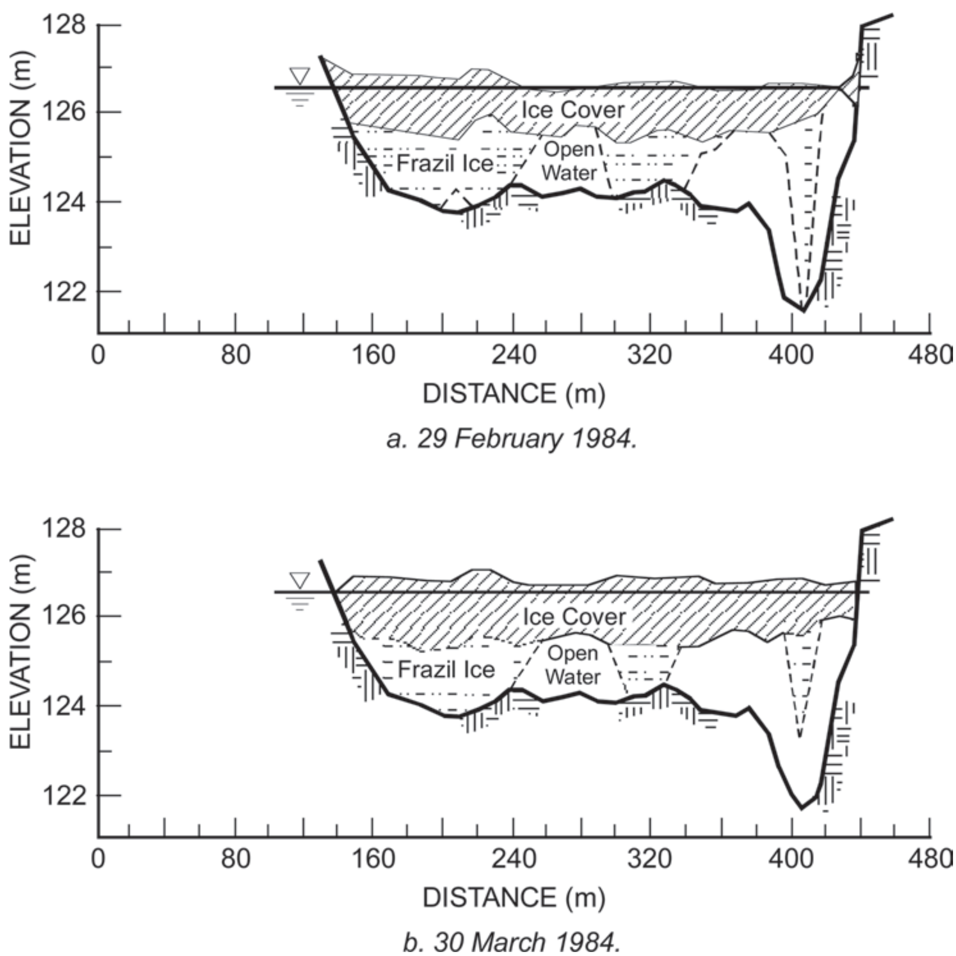


Fig. 13-10. Nonuniform ice accumulation across a section of the Tanana River, Alaska (Lawson et al. 1986).

Beltaos 1995) that ice covers are free-floating; i.e., streamwise cracks separate the floating ice cover from adjoining bank-fast ice. Actually, a cover may not always be free-floating. A stationary cover exposed to very frigid air may fuse to the channel banks. The cover then becomes constrained from freely floating up or down with changes in the flow, at least initially. Therefore, increasing flow discharge is forced partially beneath the ice cover, initially increasing flow velocities before flow erodes the bed beneath the cover. The extent to which a flow may be pressurized beneath a cover apparently has not yet been measured (the usual assumption is that covers are free-floating). An estimate would suggest that the pressure-head increase above the hydrostatic would be approximately equal to the ice thickness, the increment in water depth retained by the upstream end of the cover. Therefore, the thicker the cover the greater the pressurization possible. Eventually, the pressure would force the cover to bow upward. Also, as flow rises at the upstream end of the cover, some of it will pass over the cover. As the flow increases further, the upward pressure causes the cover to develop longitudinal cracks parallel to the banks and to float freely on the water surface. It is conjectured (e.g., Beltaos 1990) that pressure-flow conditions can only exist for a brief time, because small increases in flow suffice to cause longitudinal cracking of ice covers of thickness about 0.3 m or less; pressure flows may be more common under very thick covers. Some evidence (Zabilansky et al. 2000) suggests that scour of the channel bed may relieve the pressurized flow in alluvial channels. Very little information exists on this flow condition, especially with regard to how it may locally affect the channel bed and banks.

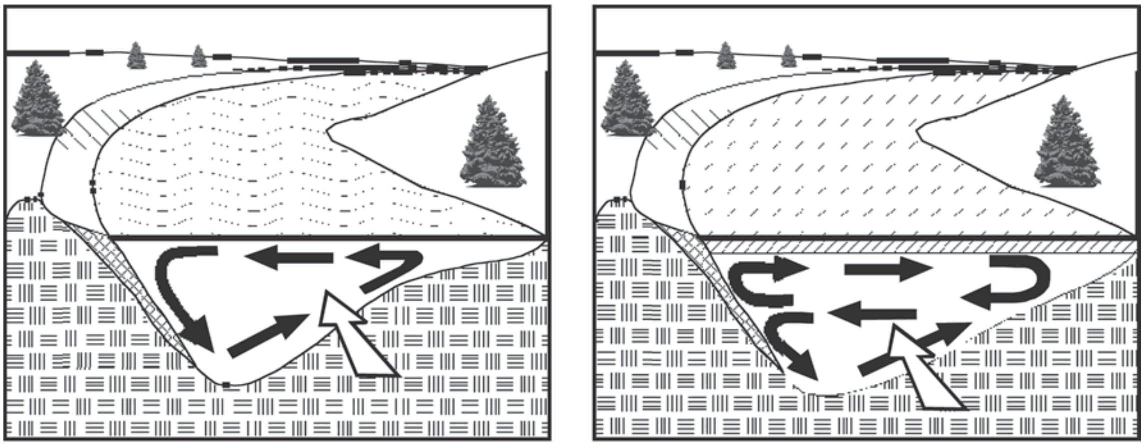
13.3.3 Ice-Cover Influence on Secondary Currents

For constant discharge, a free-floating level ice cover reduces bulk flow velocity and alters the vertical distribution of streamwise flow. In doing so, it usually dampens secondary currents.

For instance, it reduces the centrifugal acceleration exerted on flow around a river bend; though only one study has investigated this effect (Tsai and Ettema 1994). That study found that cover presence alters patterns of lateral flow distribution in a channel bend. The two sketches in Figs. 13-11 (a and b) show the main alteration, which is a splitting of the large secondary-flow spiral into two weaker spirals; owing to centrifugal acceleration acting on moving water, a large secondary-flow spiral is typical of many curved channels. The presence of a level ice cover reduces radial components of velocity and lateral bed slope in channel bends, causing the bed level to rise near the outer bank. Tsai and Ettema found a reduction in lateral bed slope of about 10%. This ice-cover effect would tend to retard bank erosion in channel bends, because it may result in reduced flow velocities near the outer bank of a bend. In other words, this effect of cover presence may dampen streamwise oscillations in bed elevation and oscillations in channel position. The dampening effect that an ice cover is calculated to have on the angle of transverse slope of the bed around a 180 bend is evident in Fig. 13-12, taken from Tsai and Ettema (1994).

13.4 ICE-COVER BREAKUP

Ice-cover breakup and clearance from a river typically coincide with substantial increases in water-flow rates in the river. Not coincidentally, these events also are periods of substantial sediment movement in alluvial rivers. For many rivers in cold regions (notably those in permafrost), breakup flows are considered to be the dominant channel-forming flows. The processes attendant to breakup and jamming are reasonably well understood; not so the impacts of breakup and jamming on channel erosion and sediment transport.



(a) (b)

Fig. 13-11. Ice-cover effects on secondary currents in a channel bend (a) open-water flow; (b) ice-covered flow.

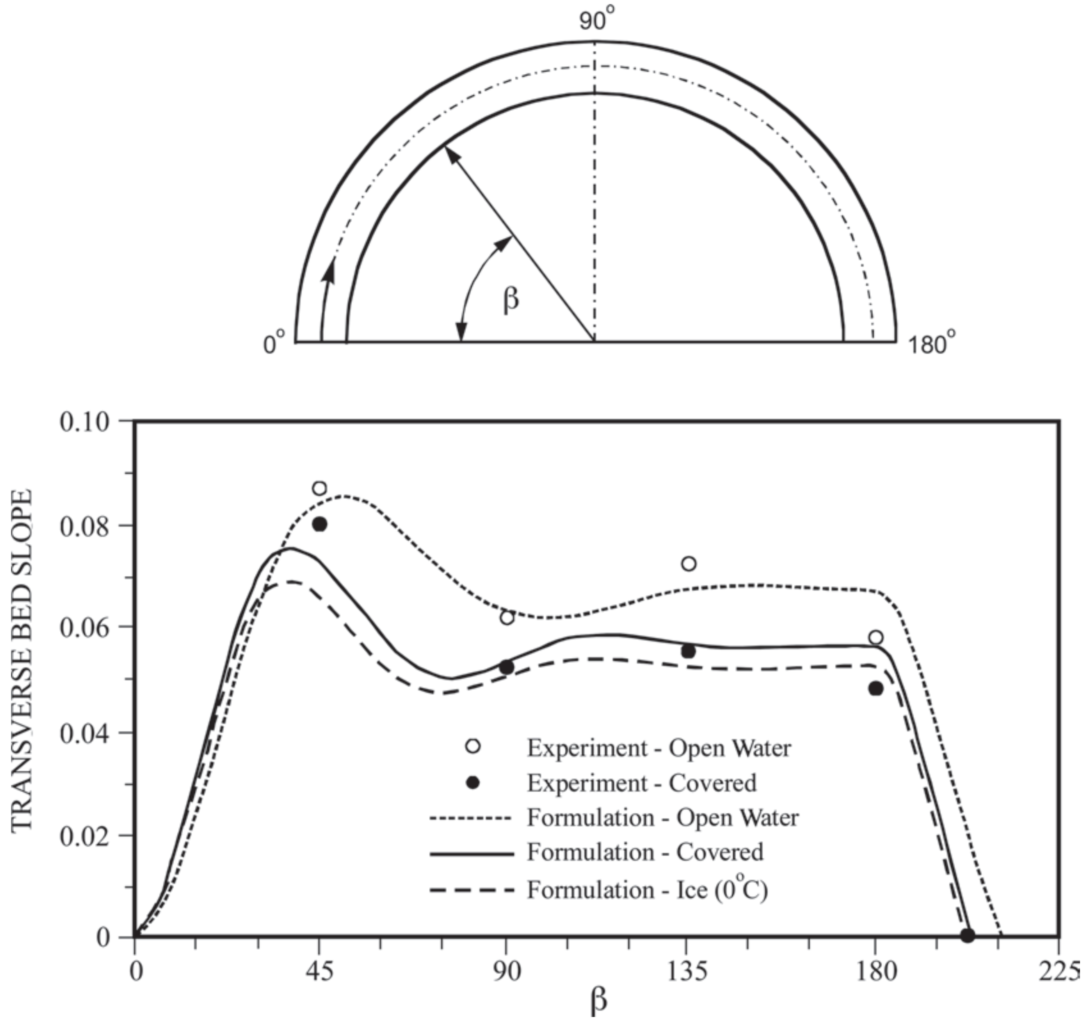


Fig. 13-12. Ice-cover effects on transverse bed slope around an alluvial-channel bend (Tsai and Ettema 1994). The effects, determined from a flume experiment and a numerical simulation, show that cover presence reduces transverse slope.

With the onset of warmer weather, ice-cover strength and thickness decrease. Ice strength usually decreases more significantly than ice thickness (Ashton 1986). In most situations, an ice cover may "rot" or "candle," becoming porous and greatly weakened before thinning. Also, with the onset of warmer weather, flow increases as snow melts, possibly accompanied by rain. Increased flow rate and depth increase the hydraulic load exerted against an ice cover, raising uplift pressure and drag, which results in hinge cracks and transverse cracks, respectively. Additionally, increased water elevation creates more surface area for ice to move.

The breakup of a river ice cover may be considered as occurring in three phases. Not all of them may occur. The phases are the prebreakup weakening of the cover, the breakup and ice run, and the breakup jam. For most river reaches, an ice cover weakens, disintegrates or breaks up, and then its fragments drift downstream. In some rivers, ice-cover breakup is followed by the development of *breakup ice jams*. For one of several reasons, certain reaches in those rivers have insufficient capacity to convey the broken ice.

The prebreakup begins with the start of runoff from the watershed when solar radiation begins to melt the snow cover, even before the average daily air temperature exceeds 0 C. The discharge in the river begins to increase, exerting an uplift pressure on the ice cover, possibly with water flowing over the cover as well as under it. With increasing discharge, the ice cover fractures in several places. For a long reach with low velocities, the break usually occurs first along the banks. The central part of the cover floats freely, but the border ice may be flooded. In areas of high flow velocity, water may rise and flow over the cover through numerous uplift fractures. Several pieces of ice may detach and begin to move downstream on the ice cover. As the discharge increases and is accompanied by daily fluctuations (responding to daytime variations in air temperature and solar radiation), ice pieces detach themselves at regions of highest flow velocity and accumulate at the front of regions of the stronger ice cover over the low-velocity reaches.

The occurrence of an *ice run* depends on a combination of flow conditions and ice-cover strength. In this regard, the direction of flow can be important. Rivers flowing into warmer regions usually begin cover breakup at the downstream end of the ice cover. The cover then progressively breaks up in an upstream direction, with the ice moving downstream in an orderly manner, provided it does not develop a jam at some congestion location. Rivers flowing into colder regions (e.g., rivers flowing north, in the Northern hemisphere) usually begin breakup near the upstream end. Breakup may also begin for river reaches for which the inflow hydrograph includes a higher peak flow rate than the outflow hydrograph, owing to flow-resistance attenuation of the hydrograph.

13.4.1 Breakup Ice Jams

It is not uncommon for ice to clear a river by means of a series of breakup jams. An initial jam forms from ice first broken over a reach upstream. Increased flow and warming cause the jam to be released, and then to be dislodged and break more ice, forming new jams downstream. Eventually, by means of this stop-go process, the flow shunts ice from the river. In the continental United States, breakup jams may occur at any time once an ice cover has formed on a river. Though spring is the usual time for breakup to occur, mid-winter thaws may cause a river to experience a series of freeze-up and break-up events. At latitudes higher than those of the continental United States, breakup jams usually occur with the onset of spring.

Many aspects of breakup-jam formation and release remain inadequately understood. An inherent difficulty with ice jams is that they radically alter the stage-discharge relationship for a river reach; a moderate flow rate in a jam-covered channel usually produces a flow stage much higher that produced by the same flow under open-water conditions. Jam formation and release may occur in fairly gentle or gradual manners. They also may occur rapidly, especially if they involve a steep hydrograph of flow or a surge. Abrupt jam release creates a surge similar to that obtained with dam-break flow; surges and ice runs have been clocked at speeds in excess of 5 m/s (Beltaos 1995).

The net effects (detrimental and beneficial) of ice-jams on channel morphology and river ecosystems have not been extensively investigated and therefore are not well understood. Of particular and common concern is the formation of ice jams at bridges, such as that shown in Fig. 13-13. Ice jams lodged against bridges not only impose substantial lateral and uplift forces on bridges, but also may aggravate constriction scour of the channel bed at the bridge site.

Fig. 13-13. Ice jam at a bridge across the Iowa River, Iowa.

13.5 SEDIMENT TRANSPORT BY ICE

Sediment-laden ice slush and clumps of ice-bonded sediment may appear during the early stages of ice formation in certain rivers and or streams subject to the winter cycle of ice formation. The ice slush and clumps comprise a mix of frazil ice and anchor ice that once was briefly bonded to the beds of such rivers and streams. The amounts of sediment entrained or rafted with the ice slush and clumps can produce a substantial momentary surge in the overall quantity of sediment moved by some rivers and streams, though at present there are no reliable measurements or estimates of ice-rafted sediment-transport rates. Much of the entrained sediment becomes included in an ice cover, where it remains stored until the cover breaks up. Though ice-rafting of sediment is known to occur (observations are reported by, for example, Barnes 1928; Wigle 1970; Michel 1972; Benson and Osterkamp 1974; Kempema et al. 1993), the implications of its occurrence largely remain unknown.

The short treatment given in this section limits itself to ice transport of sediment in rivers and streams. Shallow coastal (marine and lacustrine) waters in cold regions also are prone to bed-sediment entrainment and ice-rafting by frazil and anchor ice. Barnes et al. (1982), Osterkamp and Gosink (1983), Reimnitz and Kempema (1987), Kempema et al. (1993), Barnes et al. (1994), and Kempema (1993) describe coastal locations where ice entrains significant quantities of sediment. Storms in frigid weather conditions agitate coastal waters and can produce large quantities of frazil ice. The mechanisms whereby anchor ice forms in coastal waters include the same elements that cause anchor ice to form in rivers and streams. The formation mechanisms for coastal anchor ice are complicated, however, by the more complex flow conditions of coastal waters and by salinity considerations in marine systems. Ice can significantly affect sediment erosion and deposition in estuaries and tidal reaches of rivers. Desplanques and Bray (1986) and Morse et al. (1999) describe the influence of ice accumulation in estuaries of the northeast portion of the Bay of Fundy. The accumulations form as ice walls from stranded ice and included sediment. The ice walls confine flow and can accentuate localized channel scour. Of particular concern in this regard is scour near hydraulic structures such as bridge piers and abutments.

The mechanisms whereby ice entrains and transports sediment are not well understood. Also, the distances over which ice-rafted sediment typically may be transported are not really known. To date the only detailed laboratory investigations of the entrainment mechanisms are the studies reported by Kempema et al. (1986; 1993). A handful of experiments on anchor ice formation have been conducted, though (e.g., Tsang 1982; Kerr et al. 1998). The experiments and field observations indicate that the following two mechanisms contribute to anchor ice formation:

1. The prime mechanism is frazil ice adhesion to bed sediment. Large-scale turbulence in comparatively shallow, swift-flowing rivers and streams can mix suspended ice crystals and flocs of active frazil ice across the full depth of flow, as sketched in Fig. 13-14. When the flow is supercooled, the frazil ice may adhere to bed sediment or individual boulders and accumulate as a porous and spongy mass (Wigle 1970; Arden and Wigle 1972; Tsang 1982; Beltaos 1995). Rapids and riffles are common locations for anchor ice formation (Marcotte 1984; Terada et al. 1997). Altberg (1936) reports the occurrence of anchor ice in river flows as deep as 20 m. The foregoing references report rapid rates of anchor ice growth, such that large volumes of anchor ice form in a short period.
2. A much less significant mechanism for sediment transport is direct ice growth on the bed or on objects protruding from the bed. Together with frazil ice, supercooled water can be mixed across the flow depth. The downdraft of supercooled water chills objects in the flow (e.g., boulders and debris of various types) and enables ice to nucleate and form directly on those objects. The resultant ice crystals are relatively small and develop a fairly smooth and dense ice mass (e.g., Ashton 1986; Kerr et al. 1998).

The diurnal formation of frazil and anchor ice (as mentioned in Section 13.2) may result in repeated ice-rafting events along a river reach, each event potentially entraining substantial quantities of bed sediment. Under conditions of sufficiently frigid weather and substantial flow turbulence, extensive areas of a river's bed can become blanketed by anchor ice. Arden and Wigle (1972), for instance, describe anchor-ice formation along a several-mile reach of the Upper Niagara River, New York; the anchor ice attains sufficient bulk during a single night so that it reduces inflow into the river from Lake Erie by 20 to 30%. Usually, once the surface of a river reach is ice-covered and its water prevented from supercooling, anchor-ice formation and consequent ice rafting cease.

Because the larger sizes of sediment on a river bed protrude more into the flow, they usually are more affected by the thermal condition of the flow than by that of the bed on which they rest. Significant heat flux can occur from a sub-bed zone that is at 1 to 2°C and supercooled flow (typical supercooling is about −0.01 to −0.1 C) essentially over the full flow depth of a river or stream. Consequently, larger amounts of anchor ice typically form on coarser bed sediment. Several factors militate against extensive anchor ice formation on

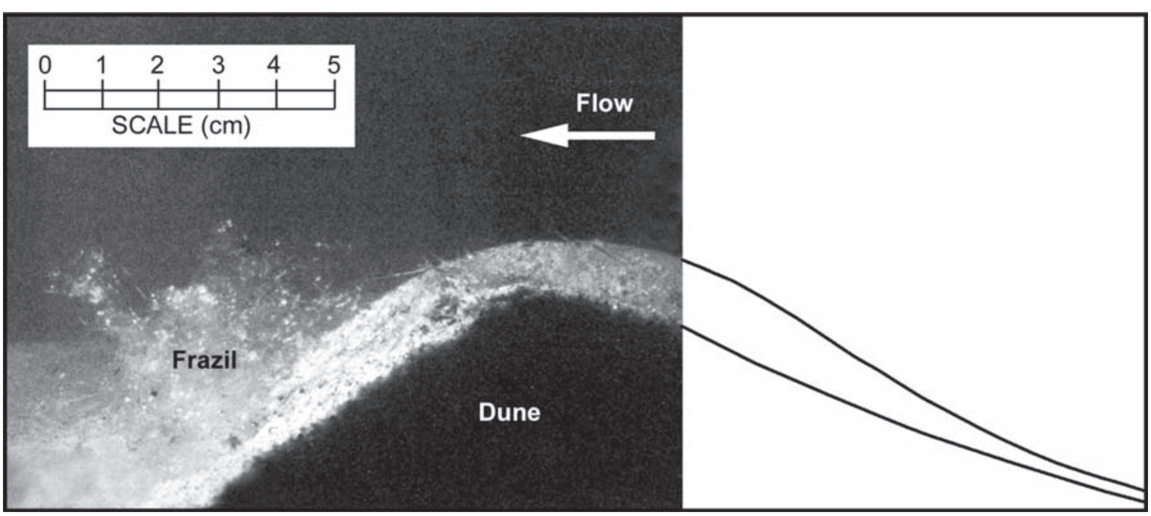


Fig. 13-14. Frazil ice accumulated in the lee of a dune, which migrates and envelops the frazil ice, eventually forming an ice-bonded clump of sand. This photo was taken from a flume experiment described by Kempema et al. (1993). (Photo taken by Ed Kempema, University of Wyoming.)

river beds of fine noncohesive sediment. In particular, such sediments are readily lifted and therefore cannot hold a significant anchor ice accumulation (Arden and Wigle 1972; Marcotte 1984).

The laboratory studies conducted by Kempema and his coworkers provide interesting insights into aspects of anchor-ice formation in the presence of bed forms. Their experiments, which were conducted with a racetrack-shaped flume fitted with sand beds in a ripple regime, show how frazil flocs become sediment-laden and lose their buoyancy as they tumble along the flume's sand bed and eventually become included within an ice-sand clump of anchor ice. As the negatively buoyant flocs of frazil and sediment accumulate in the trough of ripples (as illustrated in Fig. 13-15), they become infiltrated by sand, buried, and compressed. The resulting clumps of bonded ice and sediment may then enlarge as additional frazil flocs fuse to them, or as the clumps grow further amid supercooled water.

Eventually, a clump of anchor ice accumulation may attain sufficient buoyancy to lift sediment from the bed. The resulting concentrations of suspended sediment that the ice conveys can get quite high. Kempema et al. (1986) calculate that a neutrally buoyant clump of ice-bonded sediment may contain up to 122 grams of sediment per liter of ice and sediment. Kempema (1998) measured sediment concentrations in released anchor ice masses in southern Lake Michigan of 1.2 to 102 g/L, with an average concentration of about 26 g/L.

Accumulations of anchor ice also move gravel and cobbles. Martin (1981) mentions an instance where anchor ice entrained and moved boulders up to 30 kg in weight. Such ice rafting can move cobbles and boulders through long reaches of relatively sluggish flow deep pools in rivers.

Kempema et al. (1993) report that interactions of suspended sediment and frazil ice in the water column may directly result in the inclusion of suspended sediment in ice slush. The exact nature of the interactions and the likelihood of their occurrence, require further examination. Nonetheless, Barnes (1928) and Altberg (1936) mention an intriguing observation that frazil-ice formation appears to remove suspended sediment from a flow; after a frazil-ice event, water seems clearer. When frazil and anchor ice form, it is possible that they may diminish bed-sediment entrainment and transport. Initially, accumulating frazil and anchor ice would bind bed sediment, thereby retarding entrainment. Also, by virtue of the ice concentrations involved, frazil ice may dampen flow turbulence, a key factor in suspended-sediment transport. Once anchor ice lifts from a bed, however, it would entrain and convey sediment, although that sediment may become frozen and temporarily stored in a floating ice cover.

13. 6 ICE-COVER EFFECTS ON SEDIMENT TRANSPORT BY FLOW

The extent to which sediment transport by flow responds to ice-cover formation and presence has yet to be fully determined. Some responses are reasonably well understood, some barely recognized; few have been investigated

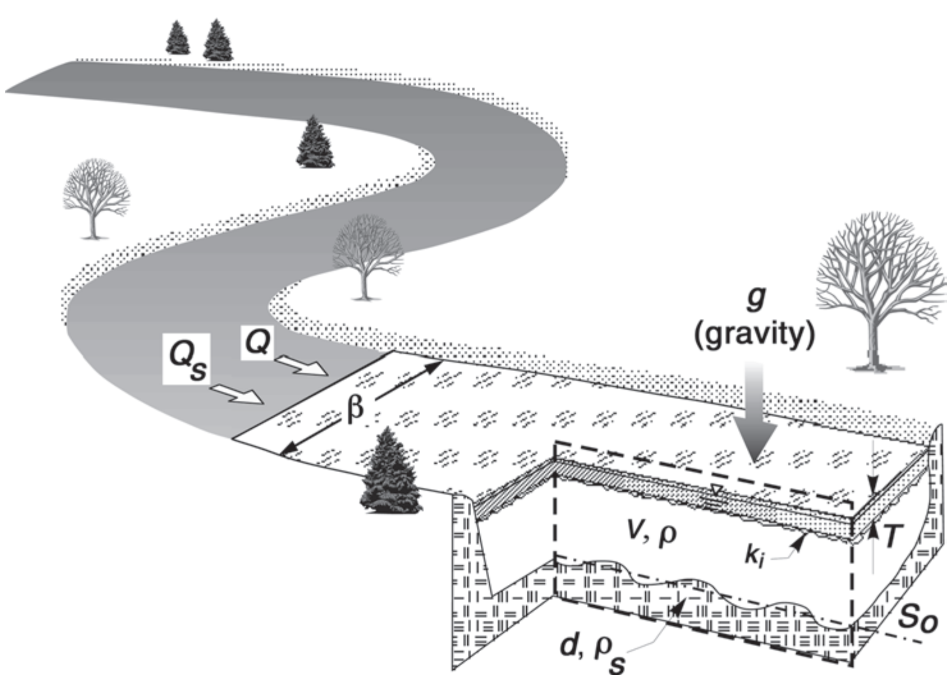


Fig. 13-15. Independent variables usually associated with flow in a loose-bed channel indicated here for an ice-covered reach.

rigorously. The interactions affect the full gamut of relationships between flow discharge and stage, macroturbulence structures, sediment-transport and mixing processes, and channel stability. This section discusses the interactions and raises practical issues stemming from them.

An essential feature of alluvial channels is that their morphology and flow-resistance characteristics alter in response to changing flow and sediment conditions. Simply put, flow and bed interact. During frigid winters, river ice modifies the interaction, over a range of scales in space and time.

The literature on river-ice hydraulics currently contains little information about ice effects on loose-bed hydraulics. Virtually all analyses of ice-covered flows (whether the ice cover is sheet ice or jammed ice) treat the bed as being fixed and thereby of constant hydraulic roughness. By the same token, the extensive literature on loose-bed hydraulics says little about flow resistance and sediment transport in alluvial channels when they are ice-covered.

13.6.1 Parameters

Dimensional analysis of variables associated with flow in a loose-bed channel (Fig. 13-15) provides a useful framework for discussing loose-bed issues in river-ice hydraulics. It quickly and formally identifies most of the interactions between cover, flow, and bed. Typically, a dependent quantity A of a channel may have the following functional dependence for flow in a reach that has a comparatively wide channel comprising a bed of uniform-diameter sediment under a uniformly thick ice cover:

$$A = f_A\left(Q, Q_S, \nu, \rho, \rho_s, d, g, S_0, B, T, k_i\right) \tag{13-5}$$

In Eq. (13.5),

Q_s = sediment discharge into reach;
ν = kinematic viscosity of water;
ρ = water density;
ρ_s = sediment density;
d = median size of bed particles;
g = gravity acceleration;
S_0 = channel slope;
B = reach width;
T = ice-cover thickness; and
k_i = hydraulic roughness of ice-cover underside.>

The dependent quantities of practical concern for the reach are flow depth, hydraulic radius, bulk velocity of flow, flow-energy gradient, sediment-transport capacity of the flow in the reach, and possibly thalweg alignment through the reach.

Though ice-cover properties T and k_i actually may also be dependent variables, especially for ice covers formed from accumulated drifting ice, here they are treated as independent variables. The variable k_i directly affects flow resistance, whereas cover thickness, T, affects flow insofar as it is of use in characterizing cover rigidity and elevation of hydraulic grade line. The present focus is on the ways in which existing ice cover modifies interactions between flow and bed. An interesting broader discussion would consider how flow and bed interaction influence cover formation. That discussion might include thermal variables, such as water temperature.

In terms of nondimensional parameters and, for convenience, considering unit discharges of water, q ($= Q/B$) and sediment q_S ($= Q_S/B$), Eq. (13-5) may be restated as

$$\Pi_A = \varphi_A\left(\frac{q}{\nu}, \frac{q_s}{\nu}, d\left(\frac{g\Delta\rho}{\rho\nu^2}\right)^{1/3}, \frac{\rho_s}{\rho}, S_0, \frac{d}{k_i}, \frac{T}{d}\right) \tag{13-6}$$

in which sediment diameter, d, is used as the scaling or normalizing length. Here, sediment discharge is total sediment discharge. Also, $\Delta\rho = \rho_S - \rho$.

The second and third parameters can be combined to express sediment transport more usefully nondimensionally as

$$\frac{q_s}{\nu}\left[d\left(\frac{g\Delta\rho}{\rho\nu^2}\right)^{1/3}\right]^{-3/2} = \frac{q_s}{\sqrt{g(\Delta\rho/\rho)d^3}} \tag{13-7}$$

For most situations, ρ_S/ρ is more or less constant (about 2.65). Thus Eq. (13-6) reduces to

$$\Pi_A = \varphi_A\left(\frac{q}{\nu}, \frac{q_s}{\sqrt{g\left(\Delta\rho/\rho\right)d^3}}, d\left(\frac{g\Delta\rho}{\rho\nu^2}\right)^{1/3}, S_0, \frac{d}{k_i}, \frac{T}{d}\right) \tag{13-8}$$

In Eqs. (13-6) and (13-8), for example, Froude number, $\mathsf{F} = (q/Y)/(gY)^{0.5}$ is a dependent parameter, because flow depth, Y, is a dependent variable.

For the case of a long, rigid, and uniformly thick, free-floating ice cover, the significance of T/d diminishes, and Eq. (13-8) simplifies to

$$\begin{aligned}\Pi_A &= \varphi_A\left(\frac{q}{\nu}, \frac{q_s}{\sqrt{g\left(\Delta\rho/\rho\right)d^3}}, d\left(\frac{g\Delta\rho}{\rho\nu^2}\right)^{1/3}, S_0, \frac{d}{k_i}\right) \\ &= \varphi_A\left(\mathsf{R}, \frac{q_s}{\sqrt{g\left(\Delta\rho/\rho\right)d^3}}, D_*, S_0, \frac{d}{k_i}\right)\end{aligned} \tag{13-9}$$

in which $D_* = d(g\Delta\rho/[\rho\nu^2])^{1/3}$, and $\mathsf{R} = q/\nu$.

Many relationships in alluvial-channel hydraulics are expressed in terms of particle Reynolds number, $\mathsf{R}_* = u_{*b}d/\nu$, and Shields parameter, $\theta = \rho u_{*b}^2/(g\Delta\rho d)$; here, u_{*b} = shear velocity associated with bed component of velocity distribution. In this regard, using $D_* = (\mathsf{R}_*)^2/\theta$, Eq. (13-9) can be recast more usefully as

$$\begin{aligned}\Pi_A &= \varphi_A\left(\mathsf{R}, \frac{q_s}{\sqrt{g(\Delta\rho/\rho)d^3}}, D_*, S_0, \frac{d}{k_i}\right)\\ &= \varphi'_A\left(\mathsf{R}, \frac{(\mathsf{R}_*)^2}{\theta}, \frac{d}{k_i}, \frac{q_s}{\sqrt{g(\Delta\rho/\rho)d^3}}, S_0\right)\end{aligned} \tag{13-10}$$

Most equations for bed load transport relate transport rate empirically to flow intensity, θ, (e.g., ASCE 1975; Raudkivi 1998). As shown subsequently in this chapter, the combined parameter $\theta d/k_i$ is convenient for indicating how cover roughness moderates θ. To simplify the discussion, the inflow and outflow rates of sediment, q_S, are taken to be equal, thereby relaxing Eq. (13-10) to

$$\Pi_A = \varphi''_A\left(\mathsf{R}, \mathsf{R}_*, \eta\frac{d}{k_i}, S_0\right) \tag{13-11}$$

which also recasts θ as $\eta = \theta/\theta_c$, thereby expressing θ relative to a critical value, θ_c, for incipient sediment movement. Many relationships for Π_A are expressed in terms of η or excess flow intensity, $\eta - 1$ (ASCE 1975).

The ensuing discussion considers how the parameters in Eq. (13-11) influence flow and sediment movement. It begins with a brief review of the pertinent cold-water properties.

13.6.2 Water-Temperature Effects

Ice is attended by cold water, usually at or slightly above 0 C. Most empirical relationships for alluvial-channel hydraulics are based on data obtained with water in the range from 10 to 20 C. All but one of the independent parameters in Eq. (13-10) directly involve water properties: ν, ρ, and $\Delta\rho$. Reduced water temperature increases kinematic viscosity, ν (it increases 100%, when water cools from 25 to 0 C), and slightly changes ρ (it increases about 0.3%, when water cools from 25 to 0 C, but attains a maximum at 4 C). An increase in ν directly reduces R and R_* values, at constant q. In so doing, it increases flow drag on the bed, decreases particle fall velocity, and thereby increases flow capacity to convey suspended sediment overall. By and large, the effect of low water temperature can be taken into account using R, R_*, and θ (insofar as it scales particle size and fall velocity relative to bed shear velocity, u_{*b}). The quantitative impacts of increased fluid viscosity on macroturbulence are unclear as yet.

A fair number of studies have investigated water-temperature effects on sediment transport or sediment fall velocity. The studies confirm that sediment-transport rate increases with decreasing water temperature. Lane et al. (1949) and Colby and Scott (1965) show such a trend in field data taken from the Missouri, Colorado, and Middle Loup Rivers. Extensive flume experiments are reported by Ho (1939), Straub (1955), Colby and Scott (1965), Taylor and Vanoni (1972), and Hong et al. (1984). Taken together, the flume data confirm that sediment transport rates increase as water temperature decreases, the increases becoming substantial when water temperature drops below about 15 C. The flume data reported by Hong et al. (1984), for instance, show that the mean concentration of bed-sediment transport increased by factors of up to 7 and 10 for a water temperature drop from 30 to 0 C. The increase, obtained with d = 0.11 mm, is attributable to increased concentration of sediment transport in a bed layer (layer thickness taken as $d\eta^{0.5}$) and increased uniformity of concentration distribution over the flow depth. The latter effect is largely owing to the reduced fall velocity of suspended particles. Hong et al. (1984) concluded that temperature reduction significantly increases bed-level concentration of sediment movement only if bed-layer Reynolds number, R_B (defined by Hong et al. as $\{u_*\, d\eta^{0.5}\}/\nu$) exceeds about 20; $\mathsf{R}_B = \mathsf{R}(\eta)^{0.5}$.

Several studies have looked at water-temperature affects on particle fall velocity (e.g., Interagency Committee 1957). No study seems yet to have looked at the settling velocity of cohesive sediments, or cohesive-sediment behavior overall, at water temperatures close to 0 C. For example, Huang (1981) examined water-temperature effects on cohesive-sediment fall velocities for the range from 32 C down only to 6.1 C.

13.6.3 Sediment Movement and Bed Forms

The overall magnitude of the tractive force (drag and lift components) that flow exerts on bed particles, together with the impacts of flow turbulence on all its scales, prescribes bed sediment motion. Ice-cover presence influences water drag on the bed and turbulence generation by redistributing flow and reducing the rate of flow energy expenditure along the bed. In so doing, cover poses three practical issues in using Eq. (13-10).

The first issue concerns estimation of τ_b or u_{*b}, shear stress or velocity associated with the channel bed. These variables are considerably more difficult to estimate than for open-water loose-bed hydraulics. A second issue is that the dependent loose-bed parameters (Π_A) of practical importance for alluvial-bed flows typically are estimated using semiempirical relationships developed for open-water conditions. Simply stated, at issue is the applicability of open-water empirical relationships to ice-covered flow.

A third issue concerns the streamwise variation of the flow and sediment-transport capacity of an ice-covered channel. If the sediment-transport capacity of an ice-covered channel is less than the rate at which sediment load is supplied to the channel, the bed must locally aggrade. If the converse holds, the bed must degrade locally. The former condition usually prevails for free-floating cover, because bulk velocity of flow decreases. The latter condition may occur when the cover is fixed and/or thick (large *T/d*), because the bulk velocity of flow under the cover increases. The various states of ice-cover condition complicate prediction of flow resistance and sediment transport in ice-covered alluvial channels.

The intrinsically complicated aspect of estimating flow resistance and sediment transport is that the single relevant length scale for ice-covered flow is the total flow depth, *Y*, which itself usually is a dependent variable. For open-water flow, flow drag on the bed can be characterized using R and D_*, because they are not explicitly dependent on flow depth and flow velocity, depending instead on q as well as water and particle properties.

Two practical concerns are whether river ice influences bed form geometry and, if so, whether its influence is describable using relationships developed for open-water flow. These issues have implications for estimation of flow resistance and mixing processes. Following from Eq. (13-10), bed form length, *L*, and steepness, δ, can be expressed functionally as

$$L_* = L/d = \varphi_{L_*}\left(\mathsf{R}, \mathsf{R}_*, \eta\frac{d}{k_i}, S_0\right) \qquad (13\text{-}12)$$

and

$$\delta = H/L = \varphi_{\delta}\left(\mathsf{R}, \mathsf{R}_*, \eta\frac{d}{k_i}, S_0\right) \qquad (13\text{-}13)$$

in which H = bed form height. Equations (13-12) and (13-13) indicate that ice-cover presence should influence bed form geometry. The practical concern is accurate estimation of η or u_{*b}. Figures 13-16 and 13-17 show that bed form geometry in ice-covered flow essentially conforms to the same relationships as prevail for open-water flow. Figure 13-17 shows additionally that an ice cover, by reducing excess flow intensity at the bed, $\eta - 1$, reduces bed form steepness for the range of values indicated.

However, there is an important cover influence not immediately evident from Eqs. (13-12) and (13-13) and Figs. 13-16 and 13-17. The influence is not adequately described in terms of cover influence on η or u_{*b}. Bed forms generate macroscale turbulence, or coherent turbulence structures. Cover presence, by redistributing flow, influences the development of macroturbulence and its consequences for bed sediment suspension as well as other dispersive processes. Recent experiments by Ettema et al. (2000) suggest that smooth level cover may invigorate macroturbulence generation, mildly increasing the frequency of structures generated from bed forms and enabling them to penetrate the full depth of flow.

13.6.4 Flow Resistance

The issues concerning flow resistance hinge on the issues mentioned above for sediment entrainment, bed forms, and macroturbulence. They entail estimation of resistance coefficients, f_b, associated with the bed and the ice cover, and then estimation of flow depth, *Y*, given *q*. From Eq. (13-11),

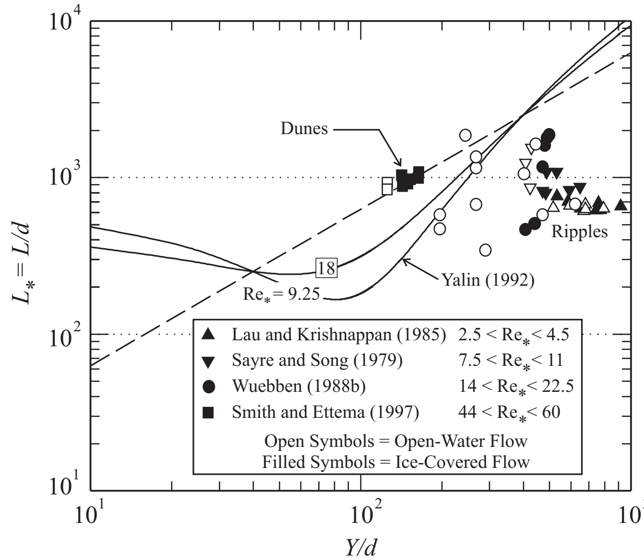


Fig. 13-16. Flume data on bedform length in ice-covered flow conform to empirical open-water curves developed by Yalin (1992).

$$Y/d = Y_* = \varphi_{Y_*}\left(\mathsf{R}, \mathsf{R}_*, \eta\frac{d}{k_i}, S_0\right) \qquad (13\text{-}14)$$

Eq. (13-14), however, is not immediately useful for predictive purposes, because open-water methods estimate *Y* as a composite of form-drag and skin-friction resistance components. It is more useful to use the Darcy-Weisbach relationship for flow in a wide channel with a free-floating ice cover written in terms of unit discharge, *q*,

$$Y_I = \left(\frac{f_I q^2}{4gS}\right)^{1/3} \qquad (13\text{-}15)$$

with flow hydraulic radius $R_I = Y/2$, $f_I = 0.5 f_b\left(1 + \frac{f_i}{f_b}\right)$, and $f_b = f_b' + f_b''$. The functional relationship for each of these component resistance coefficients can be adjusted in terms of parameters used by existing empirical, estimation relationships; e.g.,

1. for bed-surface resistance

$$f_b' = \varphi_{f_b'}\left(\mathsf{R}, D_*, S_0, \frac{d}{k_i}\right) \qquad (13\text{-}16a)$$

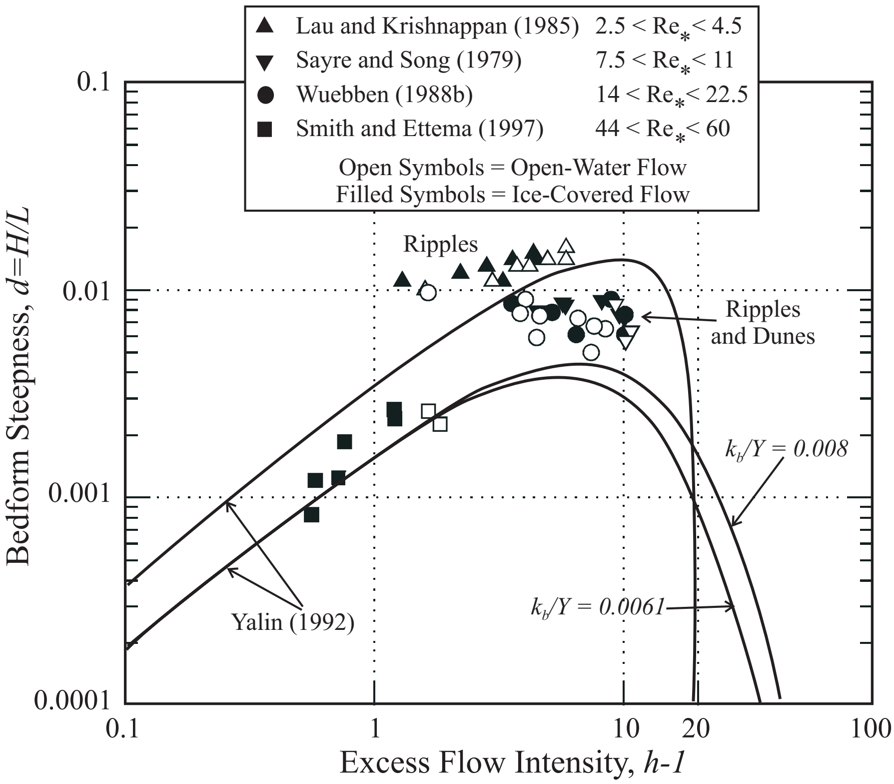


Fig. 13-17. Flume data on bedform heights in ice-covered flow conform to empirical open-water curves developed by Yalin (1992).

2. for form-drag resistance attributable to bed forms, such as dunes,

$$f_b'' = \varphi_{f_b''}\left(\mathsf{R}, D_*, S_0, \frac{d}{k_i}\right) = \varphi_{f_b''}\left(\mathsf{R}, D_*, f_b', \frac{d}{k_i}\right) \quad (13\text{-}16b)$$

3. and for the ratio

$$\begin{aligned} f_i/f_b &= \alpha = \varphi_\alpha\left(\mathsf{R}, D_*, S_0, \frac{d}{k_i}\right) \\ &= \varphi_\alpha'\left(\mathsf{R}, \mathsf{R}_{*b}, S_0, \eta\frac{d}{k_i}\right) \end{aligned} \quad (13\text{-}16c)$$

Again, an immediate practical issue implicit in Eqs. (13-16a–c) is that flow resistance in ice-covered alluvial channels can be estimated using open-water relationships, provided the influence of $\eta d/k_i$ in conjunction with the other parameters can be determined. If its influence can be determined, open-water relationships, such as those given by Einstein and Barbarossa (1952) and Engelund and Hansen (1967), can be used to predict bed resistance in ice-covered loose-bed flow. A semiempirical expression for Eq. (13-16c) is given in Fig. 13-18, which contains data from several flume studies.

Smith and Ettema (1997) developed a method, based on laboratory flume data, for estimating flow resistance in ice-covered alluvial channels. Their method is iterative and uses the following assumptions:

1. The mechanics of bed-form formation essentially is the same for open-water and ice-covered channels.
2. Methods for predicting bed-form drag in open-water flow (e.g., the Einstein-Barbarossa method or the Engelund method) can be used to predict bed form drag in ice-covered flow. This can be done by replacing the bulk drag term, $\rho g Y_I S$, with an estimate of the actual bed shear stress in an ice-covered flow.
3. The ratio of boundary shear stresses along the bed to those along the cover underside is estimated as

$$\alpha = \tau_i/\tau_b = 0.84(\eta d/k_i)^{-0.20} \quad (13\text{-}17)$$

Equation (13-17) is an equation fitted to the flume data shown in Fig. 13-18. The limits of the equation have yet to be determined for values of η beyond those indicated in Fig. 13-18.

The proposed method requires the following input variables: cover roughness, k_i; median bed-sediment diameter, d; submerged specific gravity of bed sediment, $\Delta\rho/\rho$; unit discharge of water, q; channel slope, S_0; and, an initial guess at flow depth, Y_I (say, $Y_I \approx 1.2Y_O$). The procedure uses the Einstein-Barbarossa method for predicting bed form resistance and predicts values of flow depth, Y_I.

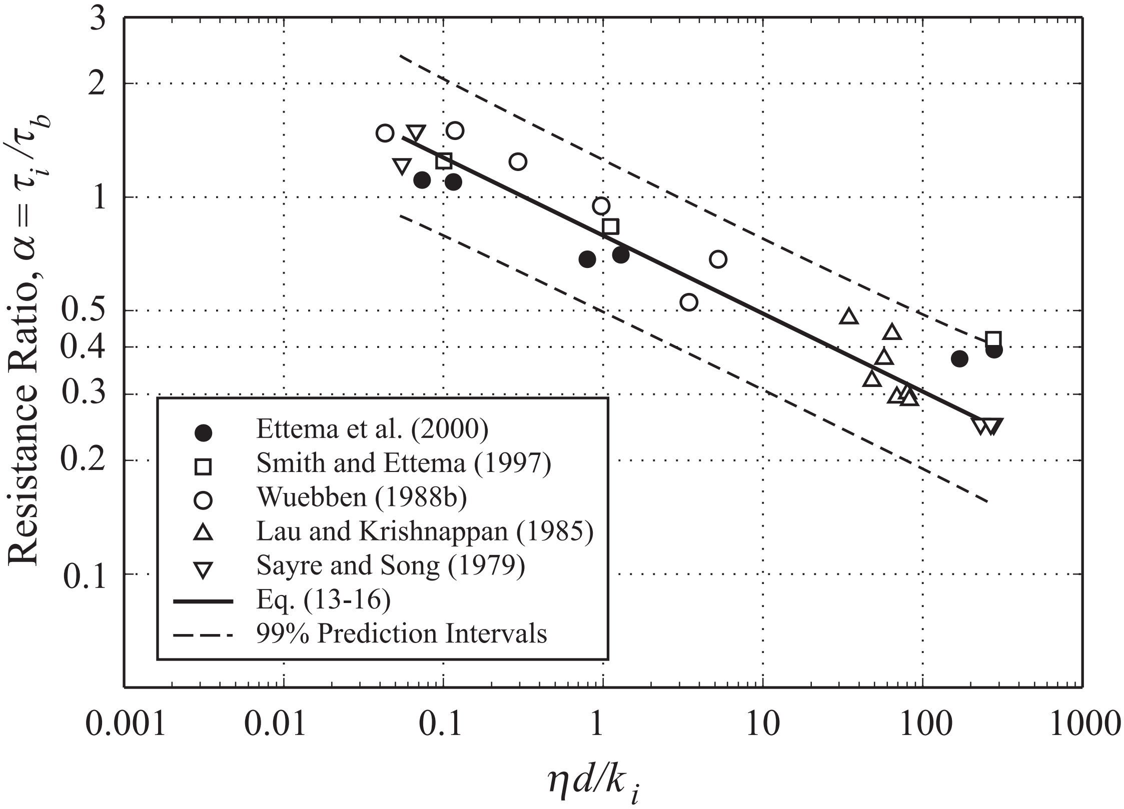


Fig. 13-18. Resistance ratio, α, for an ice-covered flow in an alluvial channel.

13.6.5 Bed-Sediment Transport

A basic issue concerns an imbalance between rate of bed-sediment supply to an ice-covered reach, q_s, and the sediment-transport capacity of that reach, q_{sI}. This issue involves the complex problem of spatially varied flow and sediment transport, with all its repercussions for local channel slope and morphology. The sediment-transport capacity of an ice-covered channel can be expressed functionally as

$$\frac{q_{SI}}{\sqrt{g\left(\Delta\rho/\rho\right)d^3}} = \varphi_{q_{SI}}\left(\mathsf{R}, \mathsf{R}_{*b}, S_0, \eta\frac{d}{k_i}\right) \qquad (13\text{-}18)$$

This equation functionally characterizes bed-load and suspended-load portions of bed-sediment transport. A fundamental issue relates directly to estimation of η or R_{*b}. However, cover influence on macroturbulence now becomes especially significant, because macroturbulence affects sediment entrainment and suspension.

13.6.5.1 Laboratory Data When examined in terms of η, or u_{*b}, data on bed load capacity of ice-covered flow experiments concur well with the open-water trend shown in Fig. 13-19 for Meyer-Peter and Mueller's formulation (1948) and Einstein's method (1950). Essentially, if η can be estimated, bed-load transport in an ice-covered channel can be estimated using an open-water method, such as the two used in Fig. 13-19. The data in Fig. 13-19 encompass the dune-bed and ripple-dune regimes.

Estimation of suspended load in an ice-covered channel is not as straightforward as bed-load estimation. Suspended load depends not only on the bed shear stress, or u_{*b}, but also on macroturbulence and flow distribution. As mentioned previously, cover presence likely significantly alters these. So far, there is no direct way to account for macroturbulence effects on suspended load.

By virtue of its reduction of bulk velocity of flow, U, and thereby τ_b and η, a free-floating ice cover typically reduces a channel's capacity to transport bed sediment. At certain zones within a channel, where the cover concentrates flow, sediment-transport rates may increase locally, however. Several laboratory studies have investigated cover-presence effects on sediment transport rate (Sayre and Song 1979; Wuebben 1986; Wuebben 1988b; Smith and Ettema 1995; Ettema et al. 2000). They all involved a free-floating cover that rises and subsides with changing flow rates. Their findings confirm that cover presence reduces rates of sediment transport. The rates decline rapidly with cover presence. Bed-load transport rate, for instance, can be almost halved

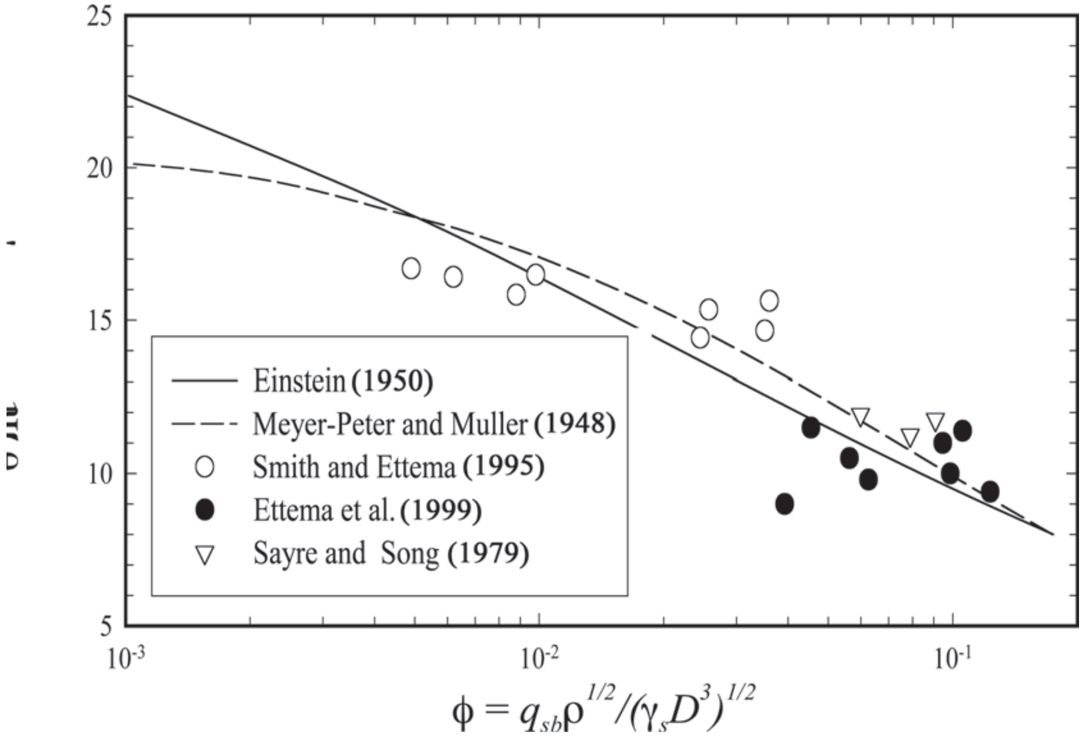


Fig. 13-19. Bedload data compared with curves generated using Einstein's procedure and the Meyer-Peter and Muller formula developed from open-water data. Covered-flow data conform to the same curves developed using open-water data.

by an ice cover that raises flow depth 15%, for a constant flow rate; this estimate assumes, reasonably, that bed-load transport rate $\propto \tau_b^2 \propto U^4$, with U decreasing by 15%; τ_b is shear stress acting on the bed and U is bulk velocity of flow (ice-covered or open water). An important point here is that sediment eroded under an ice cover may not be transported far from the erosion location.

13.6.5.2 Field Data on Bed-Sediment Load Few field studies have been conducted in which rates of sediment transport were measured for ice-covered channels. The studies indicate the inherent difficulty of obtaining such measurements and of interpreting them. Lawson et al. (1986) conducted an extensive study of flow and sediment movement at a reach of the Tanana River, Alaska. They obtained measurements of bed-load and suspended-load rates at one cross section. The rates were comparable in magnitude to rates measured during a survey conducted about a year earlier at two cross sections in close proximity to that used by Lawson et al. Burrows and Harrold (1983) describe the earlier survey. Together, these data sets indicate a great reduction in the ratio suspended load relative to the bed-load from summer to winter. The reduction is attributed tentatively to reduced flow of melt water from glaciers drained by the Tanana River. Laboratory data obtained by Lau and Krishnappan (1985) and Ettema et al. (2000) show the opposite result, which both studies attribute to cover underdamping of turbulence generated by flow over bed forms.

Alterations in flow distribution often complicate evaluation of ice-cover effects on transport rates for. This difficulty is evident in Fig. 13-2, which shows an ice cover over the Yellowstone River, near Fallon, Montana, and from figures such as Fig. 13-10, which shows nonuniform ice accumulation across the Tanana River. The series of shear lines evident in the ice cover on the Yellowstone River (Fig. 13-2) indicate that the flow area has successively narrowed. Flow-width alteration is more difficult to predict than flow depth change due to ice. The formation of subchannels within an ice-covered channel may accentuate narrowing of the flow area, especially if the channel is not prismatic. The subchannels form when accumulations of frazil slush or other ice pieces develop under the ice cover. In effect, they duct the flow in a manner that significantly alters the flow distribution from that attributable to the imposition of a level ice cover.

13.6.5.3 Field Data on Suspended Load The few field studies on sediment transport during ice-covered flow focus on suspended load and do distinguish between bed sediment and washload sediment.

The study carried out by Tywonik and Fowler (1973) focused on the measurement of suspended-sediment load in several rivers in the Canadian prairie (e.g., Assiniboine River and Red River). They report that periods of ice cover on these rivers coincide with periods of low discharge and, therefore, low rates of suspended-sediment transport. In addition, they experienced considerable difficulty in making the suspended-load measurements, owing to frigid weather conditions and the presence of slush ice.

For most cold-region rivers, the major sediment-transport event each year occurs during the large flows associated with ice runs resulting from the dynamic breakup of an ice-cover or the release of a breakup ice jam, if a jam develops. In addition to the large flow rates usually involved, these events may produce severe gouging and abrasion of banks by moving ice. The resultant sediment transport comprises a mix of bed sediment and fine sediment washed into the river during snowmelt. Bed-load measurements are very difficult to obtain under ice-run conditions.

Two studies, though, have provided some suspended-load data from individual breakup events. Prowse (1993) measured suspended-load concentrations during ice breakup of the Liard River, Northwest Territories. His data show a gradual increase in concentration with increasing water discharge immediately prior to breakup. When breakup occurred, suspended-load concentration increased by an order of magnitude, being comparable to concentrations associated with peak open-water flows of about two to five times the peak flow at breakup. Data obtained by Beltaos and Burrell (2000) during ice breakup on the St John River, N.B., show a similar trend.

13.6.6 Local Scour beneath Ice Jams

The erosive behavior of a flow may increase locally beneath an ice jam if the jam concentrates flow, increasing the magnitude of its velocity and turbulence. Also, an ice jam may deflect flow, altering its direction in a manner that aggravates bank erosion or channel shifting. This mechanism locally increases flow velocity, and it may occur when flow and ice pieces are forced beneath an ice accumulation, such as an ice jam or an ice cover. Localized scour of an alluvial bed or bank of a channel may occur in the vicinity of an ice cover when the flow field at the cover locally increases flow velocities and thereby increases flow capacity to erode bed or bank sediment. There are several conditions under which this mechanism may occur.

The most severe condition typically occurs near the toe of an ice jam (freeze-up or breakup), as illustrated in Fig. 13-20. There, where jam thickness is greatest and flow most constricted, increased flow velocities may locally scour the bed (Neill 1976, Mercer and Cooper 1977, Wuebben 1988a). Channel locations recurrently (nominally every year) subject to ice jams may develop substantial scour holes. Tietze (1961) and Newbury (1982), for example, suggest instances of such scour holes at sites of recurrent freeze-up jams. In most circumstances, the scour hole would have no lasting or adverse effect on channel morphology, because it would gradually fill once the jam was released. It is conceivable that in certain circumstances, nonetheless, the localized scour could have a longer-term effect on channel morphology—e.g., if it promoted bank erosion at the jam site, or led to the washout of the channel feature triggering the jam, such as an island or bar.

To a lesser extent, local scour of bed and bank may also occur when ice pieces collect at the leading edge of an ice cover or at some channel feature (e.g., a set of channel bars) that impedes their drift. These situations are quite marginal in extent, likely occurring more or less randomly along a channel, and are short-lived. However, they potentially may trigger more severe erosion in some situations.

13.6.7 Constriction Scour and Local Scour at Bridge Piers

The consequences of ice-cover presence for constriction scour and local scour depth at bridge piers have yet to be examined. The prime ice-related concern is that a bridge crossing may congest ice passage in a river and, thereby,

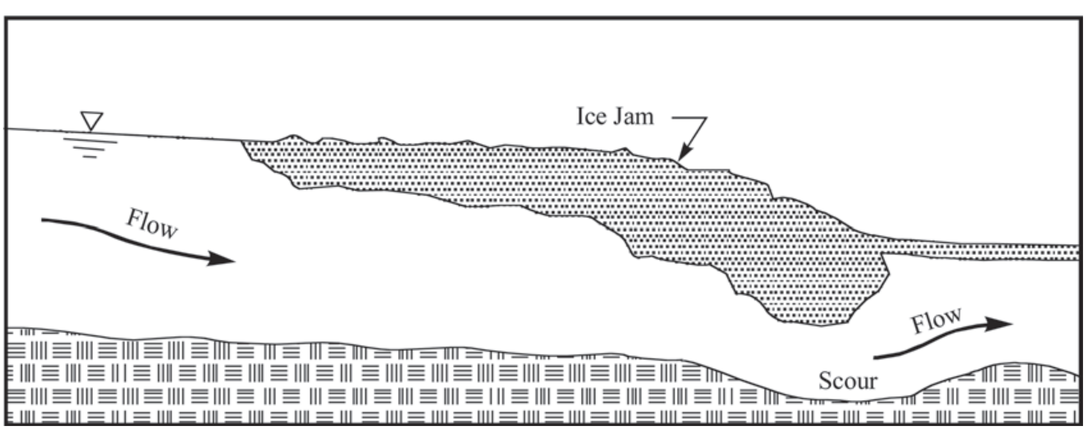


Fig. 13-20. Flow acceleration and local scour beneath an ice jam.

trigger an ice jam, such as illustrated in Fig. 13-13. Should the toe (thickest part, as sketched in Fig. 13-20) of the jam coincide with the bridge site, the jam would constrict flow through the bridge site and consequently aggravate constriction scour, as well as local scour caused by pier or abutment presence. Wuebben (1988a) describes this situation. Some work has gone in to developing instrumentation for monitoring scour depths near bridge piers in ice-covered flow (Zabilansky 1998). In summary, though, ice effects on scour likely are essentially the same as those caused by debris accumulation at bridges.

As mentioned in Section 13.5, the formation of ice walls in tidal channels can constrict flow through bridge openings in such channels (Desplanques and Bray 1986).

13.7 RIVER-ICE EFFECTS ON ALLUVIAL-CHANNEL MORPHOLOGY

An open question is the extent to which the seasonal appearance and disappearance of river ice perturbs the stability of alluvial channels in cold regions. It seems from limited field observations (e.g., Mackay et al. 1974; Zabilansky et al. 2002) that river ice may exert a compound impact of hydraulic and geomechanical impacts that continually destabilize certain planform geometries of channel subject to substantial, reservoir-regulated flow during frigid winters.

The volume of literature dealing with river-ice influences on channel morphology and bank erosion is not large. Moreover, what exists contains a fair amount of hypothesis and conjecture; there is a lack of rigorous investigation into most ice effects. Indeed, the issues of whether ice modifies channel morphology and reduces or amplifies bank erosion are still matters of considerable debate. On one hand, some articles (e.g., Neill 1982; Blench 1986) largely seem to dismiss the influences. On the other hand, there are fairly numerous anecdotal articles (e.g., Marusenko 1956, Lane 1957, Collinson 1971, MacKay et al. 1974, Hamelin 1979, USACE 1983, Doyle 1988, Uunila 1997, Milburn and Prowse 1998) and the odd review article (Ettema 1999) suggesting ways in which river ice perceptibly affects channel morphology. The dismissive articles would seem to draw their conclusions overhastily, basing them on cursory observations of overall planforms of a few rivers. They do not consider the impacts of reservoir regulation of flow during winter, take into account the diversity of channel morphologies, nor consider the important ephemeral impacts of ice that trigger local changes in thalweg, without appearing to alter channel planform appreciably. There is a need for quantitative information documenting and ranking the importance of ice impacts.

Several factors influence alluvial channel stability. Most of them are explainable in terms of the equation

$$\Pi_A = \varphi_A''\left(\mathsf{R}, \mathsf{R}_*, \eta\frac{d}{k_i}, S_0\right) \tag{13-11}$$

Dependent variables of practical interest are average depth of flow, Y, and hydraulic radius, R, as well as channel width, B; sinuosity, ζ, and shape; flow-energy gradient, S; and sediment-transport capacity, q_{sI}. Significant changes in any of the independent variables in Eq. (13-11) may alter R, ζ, or q_{sI} and may destabilize the alluvial reach. The greatest natural disturbances typically result from changes in water and sediment inflow rates q, or q_s. (In some respects, ice influences on channel morphology are discussed more conveniently in terms of total discharge rates Q and Q_S, because of the three-dimensional nature of channel morphology. The present discussion, however, continues in terms of unit discharges.)

A relatively long, level ice cover, for instance, practically doubles the wetted perimeter of flow in a channel, and it thereby significantly increases the boundary resistance exerted on the flow. Ice accumulated as an ice jam increases flow resistance by locally constricting flow. Increased flow resistance typically results in increased flow depth, altered flow distribution, and reduced flow drag on the bed—at least for fixed-bed channels. For a given channel, ice impacts on channel bed and banks increase in significance as unit water discharge, q, increases. Sediment entrainment and transport increase with increased flow in a channel when ice-covered channel as with open-water flow. Increased flow also increases the velocity of moving ice and increases the possibility of over-bank flow. River-ice impacts likely become more significant when water discharge fluctuates appreciably; then the prospects for other adverse ice influences increase, such as ice-cover breakup followed by ice jamming.

13.7.1 Hydraulic Impacts

River ice may exert the following hydraulic impacts on a channel reach:

1. By reducing the sediment-transport capacity of a river reach, ice redistributes bed sediment along the channel. Whatever local effects river ice may exert, overall river ice usually reduces the channel's overall capacity to convey the eroded sediment a significant distance from the erosion location. Consequently, bars may develop in response to flow conditions under river ice and be washed out shortly after the cover breaks up. In situations where a significant load of bed sediment enters a long reach that has a free-floating ice cover, river ice may tend to cause mild aggradation of the channel it covers. In situations where the reach is under a fixed ice cover, local degradation may occur.
2. Through its effects on lateral distribution of flow resistance and, thereby, flow and boundary drag, river ice may modify channel cross-sectional shape developed under open-water flow conditions.

3. Congestion or jamming of river ice at one channel location may divert flow into an adjoining channel, which then enlarges (anabranching/thalweg avulsion), or over a bank, which may result in a channel cutoff (avulsion).
4. Difficulties in ice passage through channel confluences may initiate ice jamming at confluences (Ettema et al. 2000; Ettema and Muste 2001). In turn, an ice jam may modify confluence bathymetry.
5. By imposing additional flow resistance, a free-floating ice cover diminishes the effective gradient of flow energy available for sediment transport and alluvial-channel shaping. It consequently alters channel-thalweg alignment.

Ice jams, especially breakup ice jams, likely exert the greatest ice-hydraulic impact on unregulated alluvial channels. Mackay et al. (1974), for instance, describe the significant impacts that breakup ice jams exert on the Mackenzie River. For channels regulated by reservoirs used for hydropower generation during winter, ice-cover formation and presence can exert significant effects (e.g., Zabilansky et al. 2002). In overall terms, ice impacts have yet to be rigorously investigated or even to be assessed quantitatively. Brief discussions of the impacts ensue.

13.7.1.1 Ice-Cover Influence on Local Elevation of Channel Bed A basic issue concerns an imbalance between unit rate of sediment supply to an ice-covered reach, q_s, and the sediment-transport capacity of that reach, q_{sI}. This issue involves the complex problem of spatially varied flow and sediment transport, with all its repercussions on local channel slope and morphology. If the sediment-transport capacity of an ice-covered channel, q_{sI}, were less than the rate at which sediment load was supplied to the channel, q_s, the bed elevation must rise locally. Conversely, if $q_{sI} > q_s$, the bed elevation must drop locally. The former condition usually would prevail for a floating cover, because bulk velocity of flow decreases. The latter condition may occur when the cover is fixed and/or thick, because the bulk velocity of flow is forced to increase substantially under the ice cover, with some flow spilling over the cover, as indicated in Fig. 13-21. An ice jam, by constricting flow, may scour a riverbed locally, especially at the jam's toe (Neill 1976; Wuebben 1988a), as shown in Figure 13-20.

13.7.1.2 Channel Anabranching, Avulsions, and Cutoffs Channels with tight meander loops or with subchannels around numerous bars or islands are prone to ice-jam formation. Such channels typically have insufficient capacity to convey the incoming amount of ice. Their morphology may be too narrow, shallow, curved, or irregular to enable drifting ice pieces to pass. Jam formation may greatly constrict flow, causing it to discharge along an alternate, less resistant course. Prowse (2001), King and Martini (1984), and Dupre and Thompson (1979) suggest that ice-jam induced avulsion plays a major role in shifting the distributary channels of river deltas. Zabilansky et al. (2002) indicate that ice-induced avulsions of subchannels may occur in sinuous-braided reaches of the Missouri River.

At sites where a river flows in two or more subchannels, ice-cover formation can trigger a switch of the principal thalweg from one subchannel to the other. Figure 13-22 illustrates the processes involved. When a rougher ice cover forms in one subchannel, the cover partially diverts flow from that subchannel to the subchannel with the smoother ice cover. The subchannel with the smoother ice cover then enlarges while the rougher-covered subchannel shrinks. Survey observations from the Fort Peck reach of the Missouri River (Zabilansky et al. 2001) suggest that thalweg switching is a recurrent process and that switches may take several winters to fully occur. Strictly speaking, such switching is a stochastic dynamic process that may be narrow-banded about a dominant period (e.g., a certain number of winters). It also may be broad-banded due to several factors (e.g., variability of flow conditions during a year or during ice-cover formation).

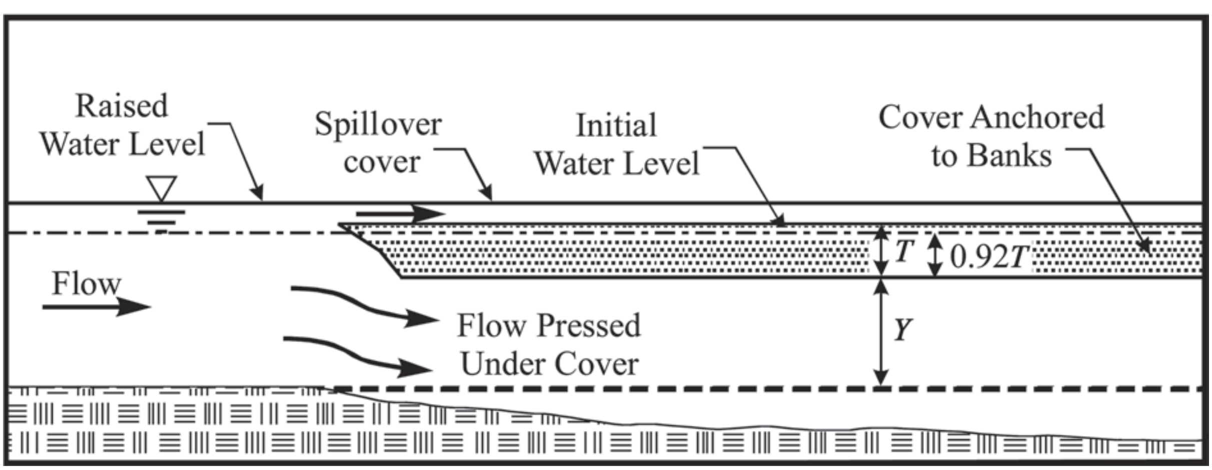
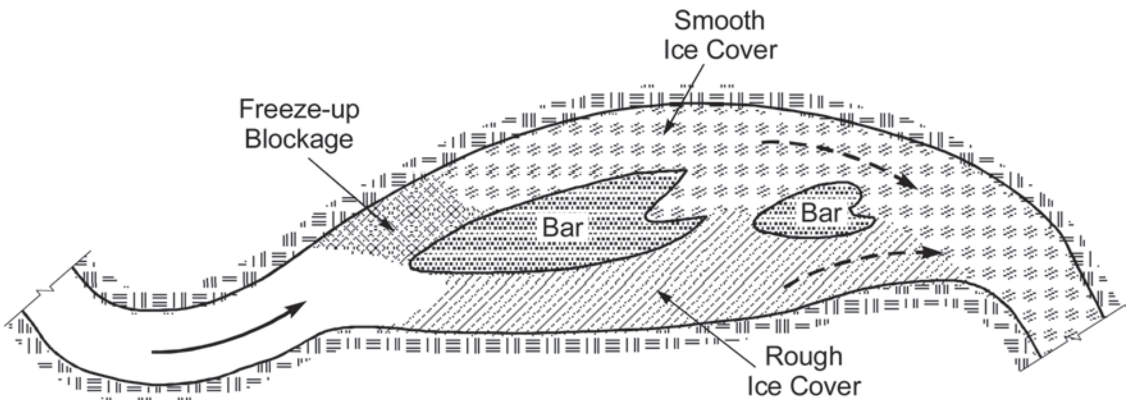


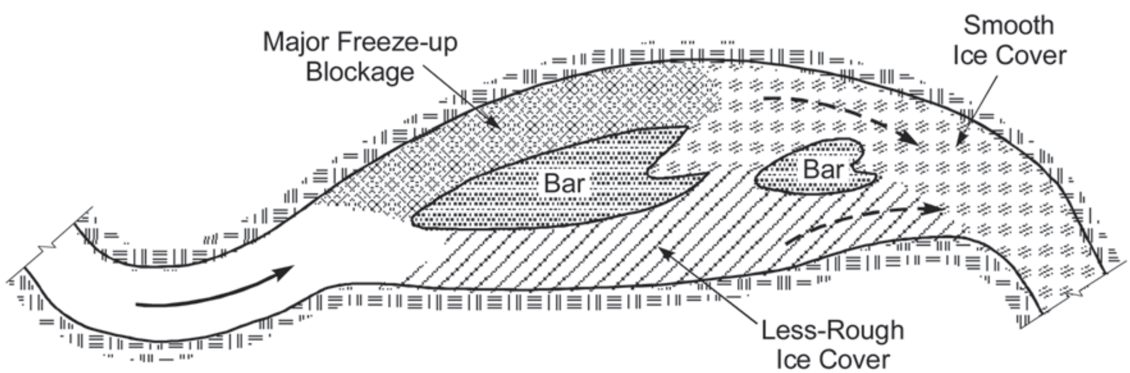
Fig. 13-21. Flow in a channel reach constricted by a fixed ice cover.

When an ice jam forms in a meander loop, upstream water levels may rise to the extent that flow proceeds overbank and across the neck of a meander loop. If the meander neck comprises readily erodible sediment and the flow is of sufficient scouring magnitude, flow diverted by the jam may result in a meander-loop neck cut, whereby a new channel forms through the neck, and the former channel is left largely cut off. A meander cutoff shortens and steepens a channel reach, the consequences of which are felt upstream and downstream of the cutoff reach. The net effect of ice jams, in this regard, is to reduce channel sinuosity. Mackay et al. (1974), for instance, cite examples of such events.

If, on the other hand, the meander loop is wide and not easily eroded, overbank flow resulting from an ice jam may have the reverse effect. Rather than the net consequence being the erosion of channel through the meander loop, overbank flow may deposit sediment, thus raising bank height and reinforcing the meander loop. Eardly (1938) reports that ice jams cause substantial sediment deposition on the flood plain of the Yukon River. A similar event is reported in Simon et al. (1999) for the Fort Peck reach of the Missouri River. Overbank deposition of sediment, together with ice-run gouging and abrasion of sediment erosion from the lower portion of a bank, may oversteepen riverbanks.



(a)



(b)

Fig. 13-22. Ice-cover formation in a sinuous-braided channel may alternate the location of the major subchannel. Two scenarios for alternation of major subchannel were identified: (a) A relatively short initial accumulation of drifting ice in subchannel 1 may divert ice into subchannel 2, which then becomes extensively enveloped by a rough ice cover. Meanwhile, subchannel 1 freezes over with a smooth ice cover, or may remain partially open. The greater flow resistance in subchannel 2 causes flow to favor subchannel 1, which then enlarges. (b) A relatively long initial accumulation of drifting ice in subchannel 1 may divert ice and flow into subchannel 2, which then becomes extensively enveloped by a less-rough ice cover. The greater flow resistance in subchannel 1 causes flow to favor subchannel 2, which then enlarges.

13.7.1.3 Channel Confluences By virtue of their role in connecting channels and thereby concentrating ice within a watershed, confluences are perceived as locations especially prone to the occurrence of ice jams. Fairly numerous accounts exist of jams in the vicinity of a confluence (Tuthill and Mamone 1997). Flow and ice concentration in a confluence may cause ice to jam within a confluent channel, within the confluence itself, or at some distance downstream from the confluence. Various mechanisms may trigger jams in the vicinity of confluences. Confluence bathymetry plays a significant role in jam initiation, and in turn jamming can modify confluence bathymetry; see Ettema and Muste (2001).

13.7.1.4 Cover Influence on Thalweg Alignment Ice cover reduces the effective energy gradient of flow (and thereby the stream power) available for sediment transport and channel shaping. Therefore, cover formation may trigger a change in thalweg alignment.

Figures 13-23(a to c) suggest that, in terms of flow drag on the channel bed, a covered flow is effectively equivalent to a deepened and slowed open-water flow. For a constant flow rate, this influence is equivalent to a reduction in channel slope (or reduced stream power). Figure 13-24 tentatively relates thalweg and channel sinuosity to channel slope (in effect, to energy gradient and stream power). It suggests that thalweg sinuosity is relatively sensitive to change in energy gradient, much more sensitive than is overall channel sinuosity. For a given flow rate, sediment provenance, and bed-sediment composition, thalweg sinuosity and channel planform change as channel slope changes. Fig. 13-24 indicates that, for a given flow rate and bed sediment size, channels lengthen or branch into subchannels as channel slope increases. Channel lengthening and branching are mechanisms whereby an alluvial-channel flow increases flow resistance (and thereby rate of energy use) to offset increased flow energy associated with a larger channel slope.

When the channel is ice-covered (Fig. 13-23b) and q is constant, flow resistance imparted by the cover deepens the flow to Y_I. The unit discharge may be written as

$$q_I = q_O = Y_I \left(8gR_I S / f_I\right)^{1/2} \tag{13-19}$$

It is assumed here that the overall reach slope, S_0, and channel width do not change significantly. Cover presence, by reducing flow velocity, reduces the portion of flow energy gradient (or stream power) expended as flow drag along the channel's bed.

For an alluvial channel, a reduction in energy gradient usually implies an adjustment in planform geometry. Because an ice cover deepens and slows flow in a channel, the channel responds as if it were at a flatter slope. In effect, the channel responds as if it were conveying an equivalent open-water flow whose cross-sectional area was as shown in Fig. 13-23(c), but whose energy gradient was reduced. The effective hydraulic radius, resistance coefficient, and energy gradient of the equivalent flow are R_e, f_e, and S_e, respectively; with $R_e \approx 2R_I$ and $S_e < S_0$. For this equivalent open-water flow,

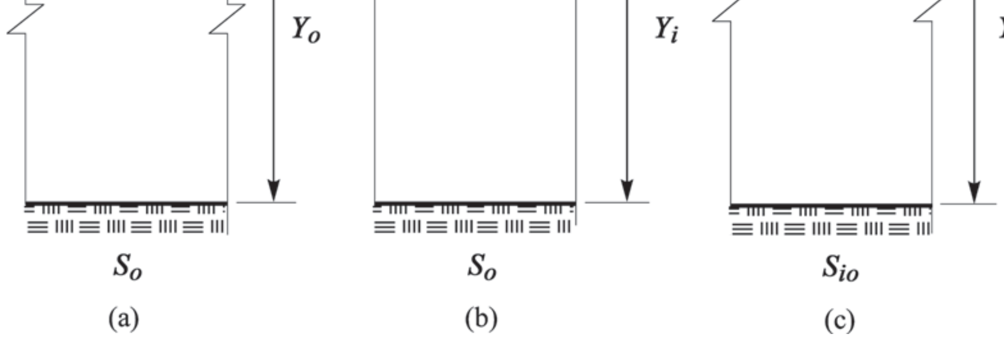


Fig. 13-23. A simplified sketch illustrating flow in an initial open-water flow (a) deepened by an ice cover (b) for the same flow rate. The ice-covered channel essentially experiences flow at a raised depth and reduced average velocity (c) (i.e., at a reduced slope, or energy gradient, S_e).

$$q_I = q_O = Y_e\left(8gR_eS_e/f_e\right)^{1/2} \tag{13-20}$$

Equations (13-19) and (13-20) give

$$\frac{S_e}{S_0} = \left(\frac{f_e}{f_O}\right)\left(\frac{Y_O}{Y_e}\right)^2\left(\frac{R_O}{R_e}\right) \approx \left(\frac{f_e}{f_O}\right)\left(\frac{Y_O}{Y_e}\right)^3 \approx \left(\frac{Y_O}{Y_e}\right)^3 \tag{13-21}$$

Equation (13-21) assumes the wide-channel approximation $R_O \approx Y_O$ and $R_e \approx Y_e$ and, because the bed sediment does not change and if the ice cover is fairly level, the ratio $f_e/f_O \approx 1$.

Equations (13-19) through (13-21), though entailing simplifying assumptions, lead to a clear result. Because covered flow depth, Y_I, usually exceeds open-water depth, Y_O, the ratio S_e/S_0 is less than 1. Therefore, the energy gradient (and stream power) available for sediment transport and channel formation decreases when a channel becomes ice-covered. For a typical situation, say, $Y_O/Y_I \approx 0.8$, $S_e/S_0 \approx 0.5$; in other words, for a given flow rate in a channel of given length, approximately half the rate of energy expenditure is available for sediment transport and channel forming. The effect of an ice cover, therefore, is to trigger a shift in thalweg sinuosity and alignment so as to balance flow-energy availability and use. However, given the magnitude and duration of flow likely needed to shift the thalweg of a channel, this ice-cover effect likely is significant only for alluvial channels whose flow is regulated by an upstream dam (notably a hydropower dam) that releases substantial flows during winter.

Figure 13-24 suggests, for instance, that halving the slope of a meandering channel (say, from 0.008% to 0.004%) will reduce thalweg sinuosity; i.e., the thalweg attempts to straighten and the meander wavelengths shorten, as sketched in Fig. 13-25.

For sinuous braided channels, as in Fig. 13-26, ice-cover formation and associated decrease in energy gradient may cause flow to concentrate in a single thalweg of greater sinuosity than the open-water thalweg. For braided channels, ice-cover presence may concentrate flow into the larger subchannels.

13.7.1.5 Jam-Collapse Surges The surge created by the collapse of an ice jam usually generates high velocities of flow that entrain considerable amounts of sediment from the channel bed as well as channel banks and possibly flood plains. As noted in Section 13.4, surge speeds up to about 5 m/s have been recorded for break-up ice jams (Beltaos 1995). Such surges can be very erosive. Anecdotal evidence exists of a case where a surge resulted in the complete removal of a small island in a river. Not unexpectedly, concentrations of suspended sediment greatly increase during the passage of a surge, as mentioned in Section 13.6.5.

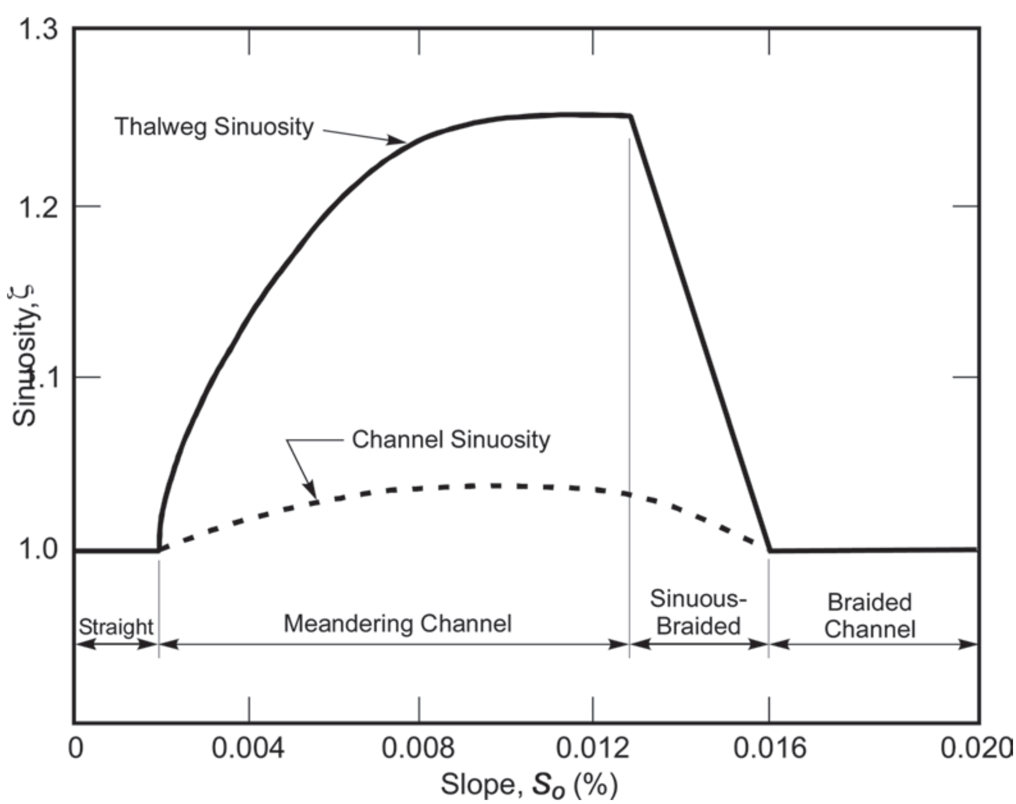


Fig. 13-24. Variation of channel and thalweg sinuosity, ζ, with channel slope, S_0. Figure adapted from Schumm and Khan (1972).

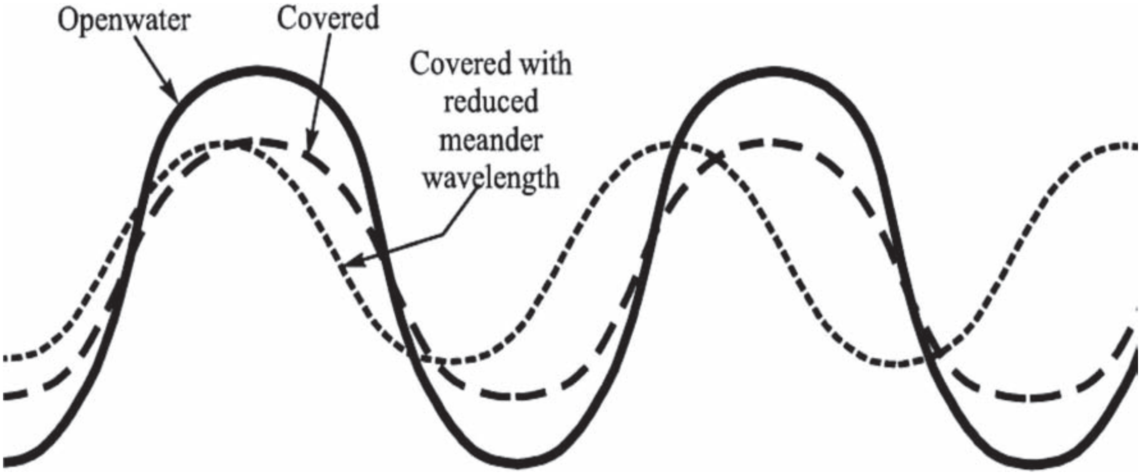


Fig. 13-25. Conceptual influence of an ice cover on a meandering channel of more-or-less uniform flow depth. The cover may cause the thalweg to straighten and meander loops to shorten.

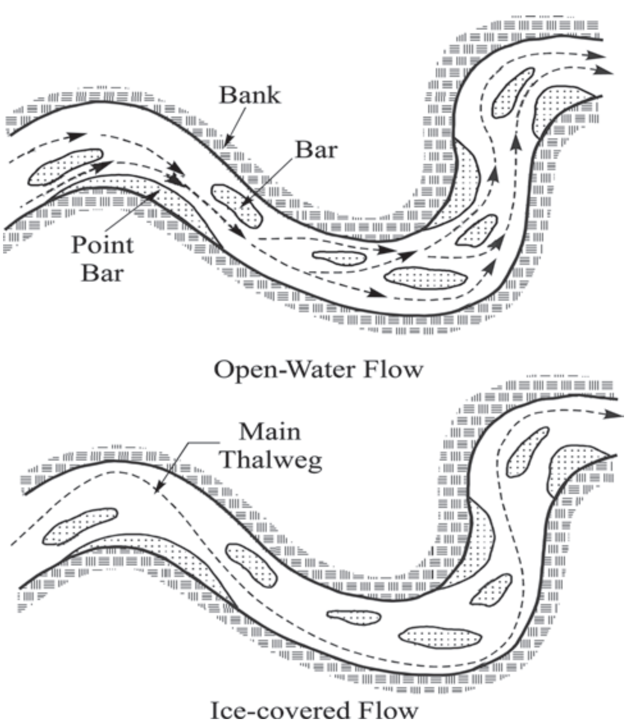


Fig. 13-26. River-ice impact on the thalweg of a sinuous-braided channel. An ice cover causes the main thalweg to become more sinuous.

13.7.2 Impacts on Riverbanks

River ice may influence channel cross-section shape, alignment, and bed elevation through several geomechanical impacts on riverbanks:

1. Reducing riverbank strength by increasing pore-water pressure or by producing rapid drawdown of the bank water table during dynamic ice-cover or ice-jam breakup. This impact is part of the overall consequences of freeze-thaw behavior for riverbanks under frigid conditions.
2. Tearing, battering, and dislodging riverbank material and vegetation during collapse of bankfast ice.
3. Gouging and abrading riverbank material and vegetation during an ice run.

The three impacts reduce riverbank resistance to scour and increase the local supply of sediment to the channel. The first two impacts are not well studied. The third has received some attention, but the extent to which it affects channel shape is unclear. It is normal for river channels and floodplains subject to ice to be denuded of larger vegetation, as is sketched in Fig. 13-27.

Engelhardt and Waren (1991), for instance, briefly describe the consequences of such combined processes for the Missouri River downstream of dams in Montana and North Dakota. Increased rates of ice-covered flow, increased movement up and down riverbanks, bank freezing at higher elevation, and more frequent freeze-thaw cycles exacerbate bank erosion. The consequences become noticeable in early spring, when large portions of riverbanks fail. Similar observations are reported by Zabilansky et al. (2001).

The ensuing subsections briefly discuss these impacts, beginning with a short review of riverbank-strength response to freezing and thawing.

13.7.2.1 Freeze-Thaw Influences on Riverbank Strength It is well known that the freezing and thawing of soil affect the erosion of riverbanks adjoining rivers and lakes. Lawson (1983; 1985) and Gatto (1988; 1995), among others, provide extensive reviews of the subject. In short, because frozen soil is more resistant to erosion

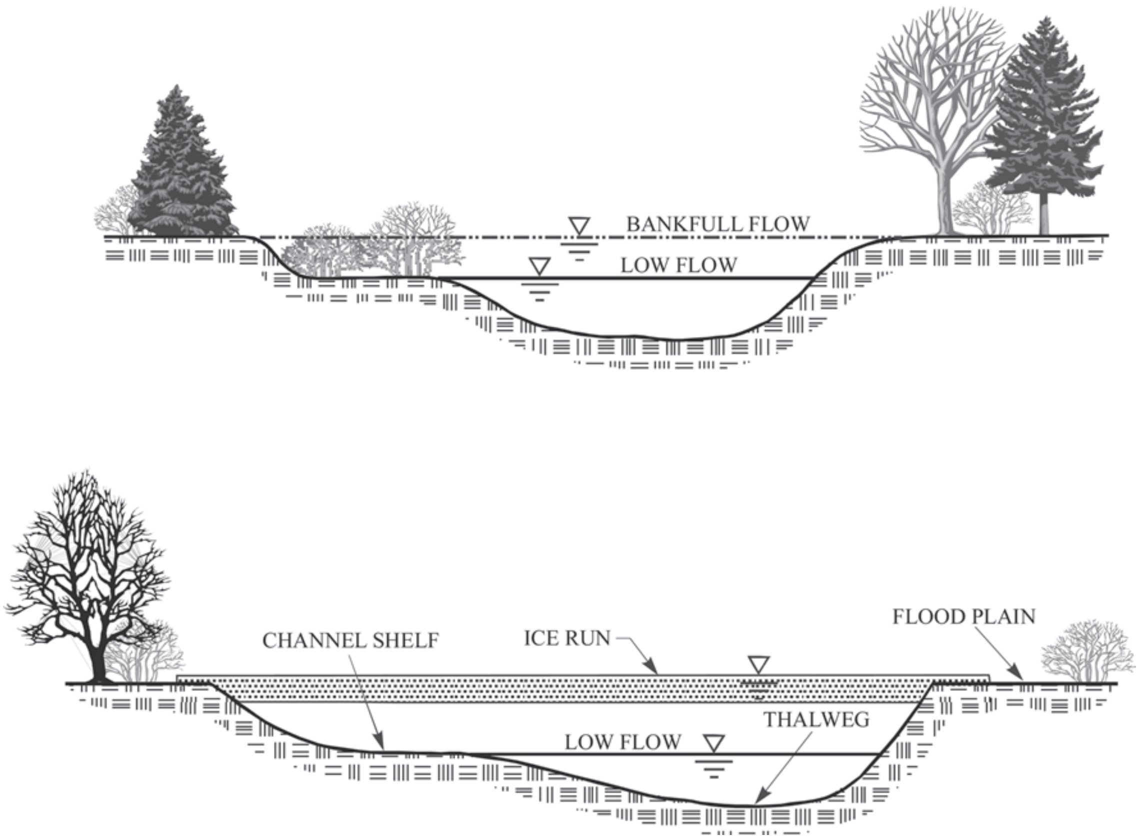


Fig. 13-27. Severe ice runs may inhibit riparian vegetation growth along riverbanks and floodplains.

than is unfrozen soil, riverbanks are less erodible while frozen. Freezing and thawing, however, usually weakens soils, making thawed (or thawing) riverbanks more susceptible to erosion. The net consequences for the overall rate of riverbank erosion, therefore, remain a matter of debate. Most likely, the net consequences vary regionally and from site to site.

Freeze-thaw cycles affect soil structure, porosity, permeability, and density. These changes in soil properties can substantially reduce soil shear strength and bearing capacity; strength reductions of as much as 95% are reported (Andersland and Anderson 1990). Such adverse effects on soil strength depend on soil-particle size and gradation, moisture content, the number and duration of freeze-thaw cycles, and several other factors. Though there is no single, standard test to determine whether a soil is prone to significant weakening due to freeze-thaw (Chamberlain 1981), particle size is commonly used as an approximate indicator of soil sensitivity to freeze-thaw weakening. Soils containing fine sands and silts are especially sensitive, because they are permeable and susceptible to change in soil structure. By virtue of their particle size (about 0.1 to 0.06 mm) and the surface-tension properties of water, fine sandy and silty soils absorb moisture more readily than do coarser or fine sediment. Clayey soils are less sensitive, because of their low permeability. The variability of soil properties along a riverbank and within a specific riverbank location causes the effects of riverbank freezing to differ along a reach.

Gatto (1995) suggests that an eroding riverbank is especially subject to deep penetration of freezing, thereby making more of the riverbank prone to freeze-thaw weakening and erosion. The absence or stunted extent of vegetation that characterizes many eroding riverbanks results in diminished insulation of the riverbank and increased heat loss to air. In addition, the crest region of a riverbank experiences greatest heat loss, owing to the crest's exposure to air on at least two sides. Because of its exposure to wind, the crest may also accumulate less snow. Less snow, in turn, means deeper frost penetration during winter and faster thaw in spring. However, less snowmelt is available to percolate into the riverbank. Questions exist about the exact manner in which border ice is anchored to the riverbank, and other factors (notably, variations in water-table (or piezometric) surface and moisture content of the top zone of the riverbank) would modify the extent of the frozen zone and its connection with river ice. Presumably, if the top portion of the riverbank and upland were dry, the riverbank crest might be the zone of least heat loss, because the distance between air and water table is greatest there.

As the upper zone of frozen ground thaws, melt water likely drains down, over the surface of the still frozen ground. The riverbank, weakened by thaw expansion of ground and subject to the seepage pressures, is in its least stable, annual condition.

Several studies (e.g., Harlan and Nixon 1978; Reid 1985) have found that south-facing riverbanks (in the northern hemisphere) experience lesser thickness of freezing, all else being equal, than north-facing riverbanks. The explanation for this is that south-facing riverbanks receive more insolation (energy in the form of short-wave radiation from the sun). South-facing riverbanks also may undergo more diurnal frequent freeze-thaw cycles (Gatto 1995). The net effect of riverbank alignment on weakening of riverbank material has yet to be determined.

13.7.2.2 Reduction of Riverbank Strength Flow stage and stage fluctuations influence seepage pressures and the freeze-thaw behavior of riverbanks. Higher flow stage raises water table in a riverbank, and a rapid drop in flow stage may momentarily reduce riverbank stability by increasing seepage pressures and thereby reducing the shearing resistance of the material comprising the riverbank. Ice-cover formation raises flow stage, whereas cover breakup may abruptly lower it. River-ice formation, thereby, may weaken riverbanks.

Riverbank freezing is closely linked to bankfast-ice formation along a channel, though the details of relationship between them are unclear. They depend on riverbank condition (material, vegetation, snow, etc.), the relative elevations of water table and flow stage, and temperatures of groundwater and river water. The strength of bankfast-ice attachment to a bank depends on the relative elevations of the water table and flow stage and on the relative water temperatures. A relatively warm (i.e., several degrees above freezing) flow of groundwater into a river will retard bankfast-ice growth and weaken its hold on the bank. The growth of a thick fringe of bankfast ice, on the other hand, may affect seepage flow through the bank, possibly constricting it and slightly raising the water table. This is especially significant for regulated rivers, for which flows do not diminish during winter.

13.7.2.3 Bankfast-Ice Loading of Bank Bankfast-ice weakening of banks likely is significant for steep banks, typically those banks containing sufficient clay to be termed cohesive. It also likely is significant for banks whose water table declines in elevation away from flow elevation in a channel, because the bankfast ice is less securely anchored into the bank. This erosion mechanism seems not to have been investigated heretofore but was observed, e.g., along the Fort Peck reach of the Missouri River (Zabilansky et al. 2001). When the flow stage in a channel drops, portions of an ice cover attached to a bank during the higher flow stage may be left momentarily cantilevered from the bank. The cantilevered ice soon collapses, weakening and wrenching bank material as it does so.

Figure 13-28 illustrates how bankfast ice might weaken a bank. The ice cover freezes into the bank. The extent of the root is limited by groundwater elevation and temperature and by the nature of the bank material. When the water level in the channel drops and the ice cover breaks up, ice attached to the bank is cantilevered out from the bank, rotates, and tears a portion of the bank as it drops. It is difficult to get direct field observations of this mechanism for bankfast ice attached to vertical banks. For the moment, evidence for

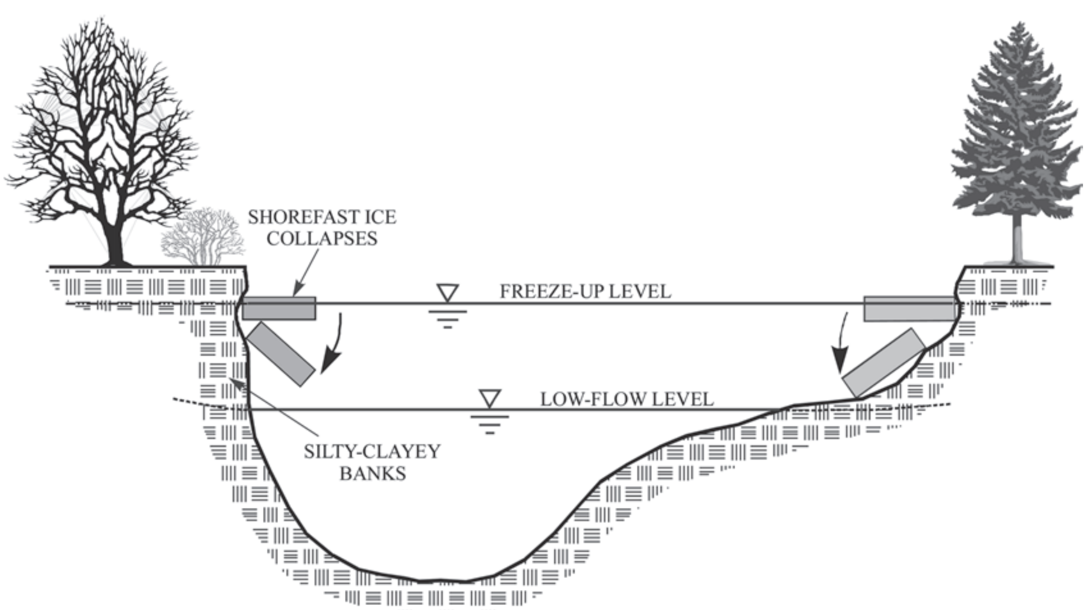


Fig. 13-28. Collapse of shorefast ice may erode banks when flow stage is lowered.

it is circumstantial. There is evidence for a related mechanism commonly termed plucking, which is the loss of riprap stones frozen to an ice sheet. Wuebben (1995), for instance, discusses plucking concerns extensively in the design of riprap for bank protection.

13.7.2.4 Gouging and Abrasion of Banks During heavy ice runs resulting from ice-cover break-up or ice-jam release, large pieces of ice potentially may gouge and abrade channel banks. There exists significant evidence showing that it substantially affects channel-bank morphology subject to dynamic ice runs (Marusenko 1956; Hamelin 1979; Smith 1979; Martinson 1980; Uunila 1997; USACE 1983; Doyle 1988; Brooks 1993; Wuebben 1995; Wuebben and Gagnon 1995). Such channels usually are relatively steep and convey high-velocity flows. Moreover, their ice covers typically break up fairly dramatically in concert with a sudden rise in flow, due, for example, to rapid snowmelt and/or rain. The resultant ice rubble comprises hard, angular blocks of ice.

One study of 24 rivers in Alberta (Smith 1979) led to the intriguing hypothesis that ice runs enlarge channel cross sections at bank-full stage by as much as 2.6 to 3 times those of comparable-flow rivers not subject to ice runs. The hypothesis is based on a comparison of the recurrence interval of bank-full flows in the 24 rivers and an empirical relationship between the cross-section area and flow rate for bank-full flow. The channel-widening effect of ice runs is plausible. However, the extent of widening indicated seems overlarge and requires further confirmation. Kellerhals and Church (1980), in a discussion of Smith (1979), argue against Smith's hypothesis. They suggest that other factors have led to an apparent widening of the channels analyzed by Smith; e.g., recent entrenchment of major rivers in Alberta and ice-jam effects of flow levels. Moreover, it is possible that the banks are somewhat protected by a band of ice forming a shear wall flanking the riverbanks. It is interesting to contrast Smith's hypothesis with a further hypothesis mentioned previously that ice jams may promote channel narrowing by causing overbank flow (e.g., Uunila 1997). For channels whose dominant channel-forming flow coincides with ice-cover breakup, overbank loss of flow reduces the flow rate to one that can be accommodated by the channel.

In many situations, notably those in which an ice run is sluggish, a shear wall of broken ice may fend moving ice from contacting the bank. The shear wall usually becomes smooth-faced, and protects riverbanks from direct ice impact or gouging. Running ice, if sufficiently thick, may still gouge the lower portion of a bank. Significant gouging may occur downstream of the toe of a jam, before the arrival of sufficient ice rubble to form shear walls. A surge front released from the jam may fracture an ice cover into large slabs, which then are set in motion. The surge front typically moves faster than the ice rubble comprising the jam, but gradually attenuates. Typically, ice gouging occurs within a relatively short reach of a river.

Ice gouging and abrasion, though, can be severe for channel features protruding into the flow. In addition, channel locations with a substantial change in channel alignment are especially prone to ice-run gouging and abrasion; e.g., a sharp bend, point bar, and portions of a channel confluence. There is a little information on how ice runs affect the local morphology of these sites. Two features have been observed in gravelly rivers: ice-push ridges and cobble pavements. Ice-push ridges form when a heavy ice run gouges and shoves sediment along the base of banks (e.g., Bird 1974). The gouged sediment piles up as ridges beneath the ice run as it comes to rest as a jam. The finer sediments eventually get washed out, leaving the more resistant gravel and boulders in ridges. The ridges usually develop in the vicinity of locations subject to recurrent ice jams.

Cobble pavements may cover bars and the lower portions of banks subject to ice gouging and abrasion. Essentially, an overriding mix of ice and cobbles removes the finer material from the surface of the bars or banks. The resultant cobble surface comprises cobbles whose major axis is aligned parallel to the channel and whose size gradually decreases downstream (Mackay and Mackay 1977). The resultant cobble pavement may extend for many miles along the banks of large northern rivers, such as the Mackenzie and Yukon Rivers (Kindle 1918; Wentworth 1932).

The gouging and abrasion of the lower portion of banks, in conjunction with overbank sediment deposition during ice-jam flooding, may produce an elevated ridge or bench feature along some northern rivers. These features have been dubbed bechevniks for Siberian rivers (Hamelin 1979). A bechevnik is the marginal strip comprising the lower portion of a riverbank and the exposed portion of the adjoining river bed that, in days gone by, formed a convenient path for towing boats upstream manually or by horse; becheva apparently is Russian for towrope. Figure 13-29 illustrates the main features of a bechevnik, which may form partly from ice abrasion and partly from the deposition of sediment and debris left by the melting of ice rubble stranded after ice runs.

Moving ice also may grind banks formed of soft rock (e.g., sandstones and mudstones) or stiff clay. Danilov (1972) and Dionne (1974), for instance, describe how moving ice has affected rock banks of rivers such as the St Lawrence River. The extent of erosion, though, is less than for banks formed of alluvial sediment.

Ice-run gouging and abrasion have an important, though as yet not quantified, effect on riparian vegetation that, in turn, may affect bank erosion and channel shifting. Where ice runs occur with about annual frequency, riparian vegetation communities have difficulty getting established. Ice abrasion and ice-jam flooding may suppress certain vegetation types along banks, as illustrated in Figs. 13-27 and 13-28 for a bechevnik, possibly exacerbating bank susceptibility to erosion. This aspect of river ice has yet to be further investigated.

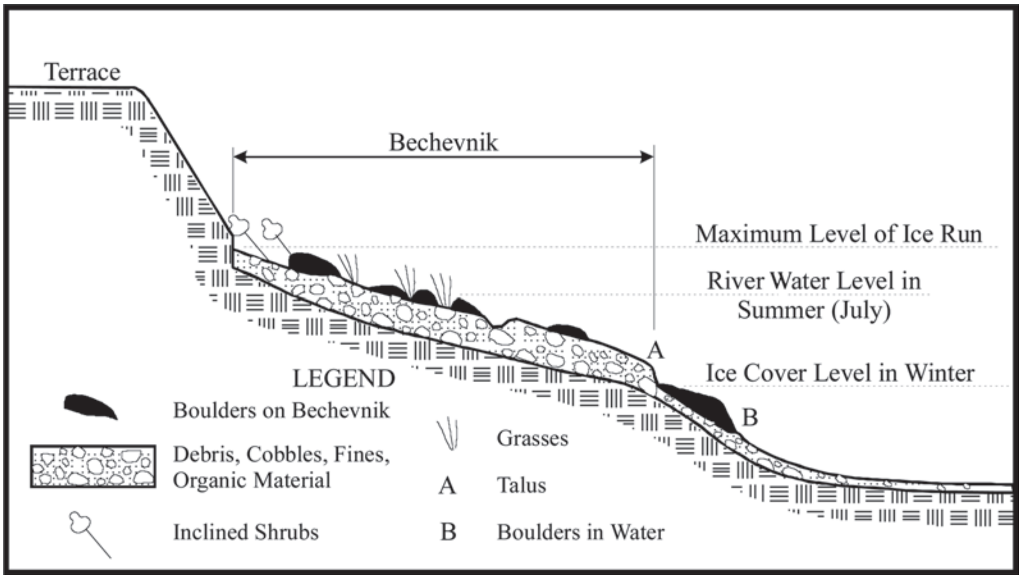


Fig. 13-29. Sketch of a bechevnik. Figure adapted from Hamelin (1979).

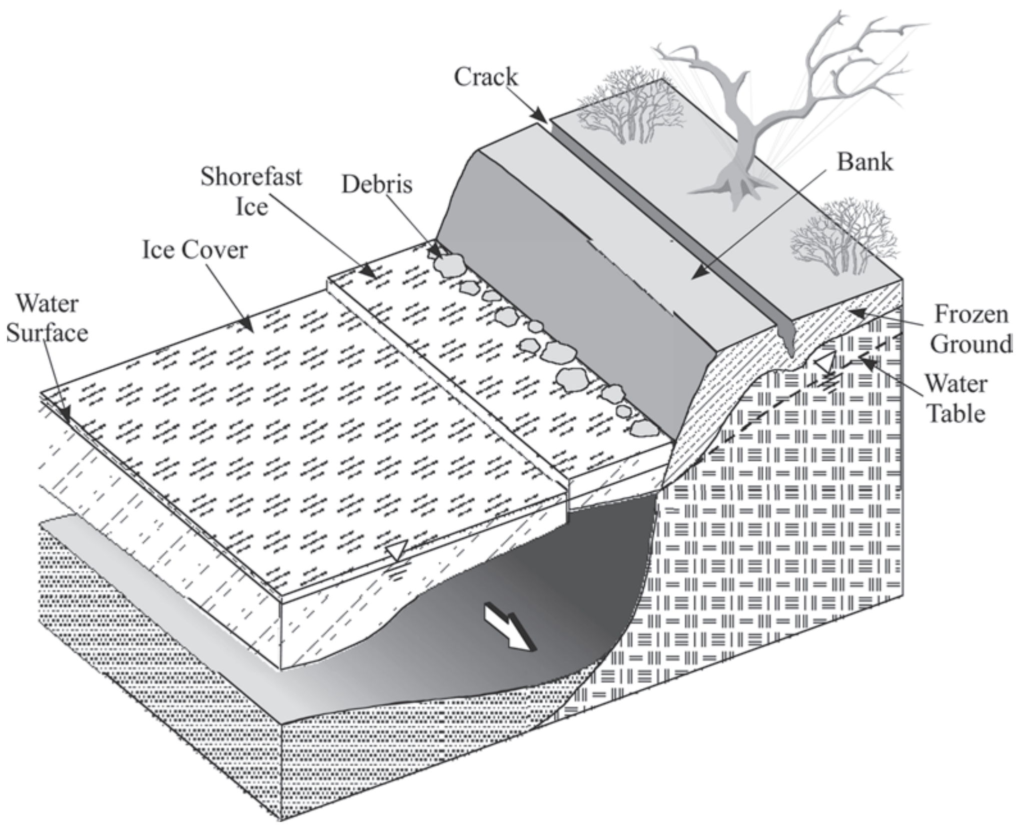


Fig. 13-30. Hydraulic impacts (e.g., thalweg shift and bank-toe erosion), together with geomechanic impacts (e.g., freeze-thaw weakening of bank material, elevated seepage pressures, bankfast-ice loading) may weaken and erode channel banks, especially along channel bends, and results in continual overall channel destabilization.

Scrimgeour et al. (1994) and Prowse (2001) provide useful early reviews.

13.7.3 Combined Hydraulic and Geomechanical Impacts on Channels

A single hydraulic or geomechanical impact of river ice may disturb a channel, but not necessarily destabilize it. A combination of hydraulic and geomechanical impacts, though, may destabilize a channel. A shift in thalweg alignment or a bank failure alone may not destabilize a channel. The channel may adjust back more or less to its stable open-water condition once open-water conditions resume. Besides, a single ice impact may be damped or possibly constrained. For instance, flow concentration along a thalweg may be damped by an increase in bed resistance resulting from an increase in bed-form size, and bank erosion may be damped as bank slope consequently flattens. High banks, which deposit a large mass of sediment into the channel, or scour-resistant strata (e.g., a clay layer or rock outcrop) may constrain thalweg shifting or entrenchment.

It probably is not surprising that channels usually considered less stable under open-water conditions are more likely to be adversely impacted by river ice. Sinuous point-bar, sinuous braided, and braided alluvial channels are especially prone to river ice impact, especially if they have steep banks formed of fine and partially cohesive sediments. The thalwegs of such channels usually lie close to the outer banks of bends, and the banks themselves are prone to bank-fast-ice loading, lack of vegetation cover (typical of eroding banks), and freeze-thaw weakening. Figure 13-30 illustrates this susceptibility. The thalweg lies close to the bank, so that the flow continually erodes the bank-toe, thereby keeping the bank steep and possibly undercutting it. Snow cannot protectively blanket the bank face. Frost penetration potentially is deep, the water table is held relatively high, and the channel shifts, destabilized.

An intriguing question is whether the destabilizing impacts of river ice uniquely modify alluvial-channel morphology. Only a tentative answer can be suggested at this moment. It is likely that the major geometric parameters do not change appreciably (e.g., channel thalweg sinuosity, width, hydraulic radius, meander radius). However, river ice likely increases irregularities in channel planform and the frequencies with which channel cross section and thalweg alignment shift. In a sense, it adds noise to the signal form of an alluvial-channel in dynamic equilibrium.

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REFERENCES

Altberg, V. I. (1936). "Twenty years of work in the domain of underwater ice formation (1915–1935)." *Bulletin, International Association for Scientific Hydrology*, 23, 373–407.

Andersland, O., and Anderson, D. M. (Eds.). (1990). *Geotechnical engineering for cold regions*. McGraw-Hill, New York.

Arden, R. S., and Wigle, T. E. (1972). "Dynamics of ice formation in the Upper Niagara River." *The Role of Ice and Snow in Hydrology*. Publication 107, International Association for Scientific Hydrology, Grenoble, France, 1296–1312.

ASCE. (1975). *Sedimentation manual.* ASCE, New York.

Ashton, G. (1986). *River and lake ice engineering*. Water Resources Publications, Littleton, Colo.

Barnes, H. T. (1928). *Ice engineering*. Renouf Publishing Company, Montreal, Que., Canada.

Barnes, P. W., Kempema, E. W., Reimnitz, E., and McCormick, M. (1994). "The influence of ice on southern Lake Michigan coastal erosion." *Journal of Great Lakes Research,* 20, 179–195.

Barnes, P. W., Reimnitz, E., and Fox, D. (1982). "Ice rafting of fine-grained sediment, a sorting and transport mechanism, Beaufort Sea, Alaska." *Journal of Sedimentary Petrology*, 52, 493–502.

Beltaos, S. (1990). "Fracture and breakup of river ice cover." *Canadian Journal of Civil Engineering*, 17(2), 173–183.

Beltaos, S. (1995). *River ice jams*. Water Resources Publications, Highlands Ranch, Colo.

Beltaos, S., and Burrel, B. (2000). "Suspended sediment concentration in the St John River during ice breakup." *Proc. Canadian Society of Civil Engineers Annual Conference*, Canadian Society of Civil Engineers, Toronto, Ont., 235–242.

Beltaos, S., and Dean, A. (1981). "Field Investigations of a hanging ice dam." *Ice Symposium*, Vol. II, International Association of Hydraulic Research, Delft, the Netherlands, 485–449.

Benson, C. S., and Osterkamp, T. E. (1974). "Underwater ice formation in rivers as a vehicle for sediment transport." *Oceanography of the Bering Sea*. Institute of Marine Science, University of Alaska, Fairbanks, Alaska, 401–402.

Bird, J. B. (1974). "Chapter 12: Geomorphic processes in the Arctic." *Arctic and alpine environments*, J. D. Ives and R. G. Barry, eds., Methuen, London, 703–720.

Blench, T. (1986). *Mechanics of plains rivers*. University of Alberta, Edmonton, Alta., Canada.

Bogordskii, V. V., and Gavrola, V. P. (1980). *Ice. Physical properties, modern methods of glaciology* (in Russian). Gidrometeoizdat, St Petersburg, Russia.

Brooks, G. R. (1993). "Characteristics of an ice-scoured river bank near Keele River confluence, Mackenzie Valley, Northwest Territories. Current Research, Part B." *Paper 93-1*, Geological Survey of Canada, Ottawa, Ont., Canada, 21–27.

Burrows, R. L., and Harrold, P. E. (1983). "Sediment transport in the Tanana River near Fairbanks, Alaska (1980–1981)." *Water Resources Investigations Report 83-4064*, U.S. Geological Survey, Reston, Va.

Chacho, E. F., Lawson, D. E., and Brockett, B. E. (1986). "Frazil ice pebbles: Frazil ice aggregates in the Tanana River near Fairbanks, Alaska." *Proc. IAHR Symposium on Ice*, International Association for Hydraulic Research, Delft, the Netherlands, 475–484.

Chamberlain, E. J. (1981). "Frost susceptibility of soil: review of index tests." *Monograph 81-2*, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, N.H.

Colby, B. R., and Scott, C. H. (1965). "Effects of temperature on discharge of bed material." *Professional Paper 462-G*, U.S. Geological Survey, U.S. Government Printing Office, Washington, D.C.

Collinson, J. D. (1971). "Some effects of ice on a river bed." *Journal of Sedimentary Petrology*, 41(2), 557–564.

Danilov, I. D. (1972). "Ice as a factor of relief formation and sedimentation." *Problemy Krioliltogii*, 2, 137–143.

Desplanqes, C., and Bray, D. I. (1986). "Winter ice regime in tidal estuaries of the northeastern portion of the Bay of Fundy, New Brunswick." *Canadian Journal of Civil Engineering*, 13(2), 130–139.

Dionne, C.-J. (1974). How ice shapes the St Lawrence. *Canadian Geographical Journal*, 88(2), 4–9.

Doyle, P. F. (1988). "Damage from a sudden river ice breakup." *Canadian Journal of Civil Engineering*, 15, 609–615.

Dupre, W. R., and Thompson, R. (1979). The Yukon Delta: a model for deltaic sedimentation in an ice-dominated environment. 11th Annual Offshore Technology Conference, 657–664.

Eardley, A. J. (1938). Yukon channel shifting. Bulletin of the Geological Society of America, Vol. 49, 343–358.

Einstein, H. A. (1950). The bedload function for sediment transport in open channels. Technical Bulletin 1026, U.S. Dept of Agriculture, Washington, D.C.

Einstein, H. A., and Barbarossa, N. (1952). "River channel roughness." *Transactions of the ASCE*, 117(2528).

Engelhardt, B., and Waren, G. (1991). "Upper Missouri River bank erosion." Report prepared by North Dakota and Montana, North Dakota State Water Commission, Bismark, N.D.

Engelund, F., and Hansen, E. (1967). "A monograph on sediment transport in alluvial streams." *Teknisk Forlag*, Technical University of Denmark, Copenhagen.

Ettema, R. (1999). "A review of river-ice impacts on alluvial-channel stability." *Canadian Workshop on River Ice*, Canadian National Research Council, Ottawa, Ont., 125–140.

Ettema, R., Braileanu, F., and Muste, M. (2000). "A method for estimating sediment transport in ice-covered channels." *Journal of Cold Regions Engineering,* 14(2), 130–145.

Ettema, R., and Muste, M. (2001). "Laboratory observations of ice jams in channel confluences." *Journal of Cold Regions Engineering,* 15(1), 41–51.

Ettema, R., Muste, M., and Kruger, A. (1999). "Ice jams at river confluences." *CRREL Report 99-6,* U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, N.H.

Gatto, L. W. (1988). "Techniques for measuring reservoir bank erosion." *Special report 88-3*, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, N.H.

Gatto, L. W. (1995). "Soil freeze-thaw effects on bank erodibility and stability." *Special Report 95-24*, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, N.H.

Gogus, M., and Tatinclaux, J-C. (1981). "Mean characteristics of asymmetric flows: application to flow below ice jams." *Canadian Journal of Civil Engineering*, 8(3), 342–350.

Hamelin, L.-E. (1979). "The bechevnik: A river bank feature from Siberia." *The Musk Ox*, 25, 70–72.

Hanjalic, K., and Launder, B. E. (1972). "Fully developed asymmetric flow in a plane channel." *Journal of Fluid Mechanics*, 51, 301–335.

Harlan, R. L., and Nixon, J. F. (1978). "Section 3: Ground thermal regime." *Geotechnical engineering for cold regions*, O. Andersland and D. M. Anderson, eds., McGraw-Hill, New York.

Ho, P. Y. (1939). "Dependence of bedload movement on grain shape and water temperature." *Mitteilungen der Preusssischen Versuchsanstalt fuer Wasser-, Erd-, und Schiffbau*, 39.

Hong, R.-J., Karim, F., and Kennedy, J. F. (1984). "Low-temperature effects on flow in sand-bed streams." *ASCE Journal of Hydraulic Engineering*, 110(2), 109–125.

Huang, J. (1981). "Experimental study of settling properties of cohesive sediment in still water." *Journal of Sediment Research*, (2), 30–42.

Interagency Committee. (1957). "Some fundamentals of particle size analysis, a study of methods used in measurement and analysis of sediment loads in streams." *Report 12*, Subcommittee on Sedimentation, Interagency Committee on Water Resources, St Anthony Falls Hydraulic Laboratory, Minneapolis, Minn.

Johnston, G. H. (Ed.). (1981). *Permafrost: Engineering design and construction*. Wiley, New York.

Kellerhals, R., and Church, M. (1980). "Comment on 'Effects of channel enlargement by river ice processes on bankfull discharge in Alberta, Canada,' by Smith, D. G." *Water Resources Research*, 16(6), 1131–1134.

Kempema, E. W. (1998). "Nearshore ice formation and sediment transport in southern Lake Michigan. Ph.D. thesis, University of Washington, Seattle.

Kempema, E. W., Reimnitz, E., Clayton, J. R., and Payne, J. R. (1993). "Interactions of frazil and anchor ice with sedimentary particles in a flume." *Cold Regions Science and Technology*, 21, 137–149.

Kempema, E. W., Reimnitz, E., and Hunter, R. E. (1986). "Flume studies and field observations of the interaction of frazil ice and anchor ice with sediments." *Open-File Report 86-515*, U.S. Geological Survey, Menlo Park, Calif.

Kerr, D. J., Shen, H.T., and Daly, S. F. (1998). "Evolution and hydraulic resistance of anchor ice on gravel beds." *Proc. of the International Association for Hydraulic Research, 14th Ice Symposium*, International Association for Hydraulic Research, Delft, the Netherlands, 703–710.

Kindle, E. M. (1918). "Notes of sedimentation in the Mackenzie River Basin." *Journal of Geology*, 26, 341–360.

King, W. A. and Martini, I. P. (1984). "Morphology and recent sedimentations of the anastomizing reaches of the Attawapiskat River, James Bay, Ontario, Canada." *Sedimentary Geology*, 37(4), 295–320.

Lane, E. W. (1957). "A study of the shape of channels formed by natural streams flowing in erodible material." *Missouri River Series Report No. 9*, U.S. Army Corps of Engineers, Missouri River Division, Omaha, Neb.

Lane, E. W., Carlson, E. J., and Hanson, O. S. (1949). "Low temperature increase in sediment transportation in Colorado River." *Civil Engineer*, 19, 619–621.

Lau, Y. L., and Krishnappan, B. G. (1985). "Sediment transport under ice cover." *Journal of Hydraulics Division, ASCE*, 111(6), 934–950.

Lawson, D. E. (1983). "Erosion of perennially frozen streambanks." *CRREL Report 83-29*, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, N.H.

Lawson, D. E. (1985). "Erosion of northern reservoir shores: Analysis and application of pertinent literature." *Monograph 85-1*, U.S. Army Cold Regions Research and Engineering Laboratory, Hanover, N.H.

Lawson, D. E., Chacho, E. F., Brockett, B. E., Wuebben, J. L., Collins, C. M., Arcone, S. A., and Delaney, A. J. (1986). "Morphology, hydraulics and sediment transport of an ice-covered river: Field techniques and initial data." *Report 86-11*, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, N.H.

Mackay, J. R., and Mackay, D. K. (1977). "The stability of ice-push features, Mackenzie River, Canada." *Canadian Journal of Earth Sciences*, 14(10), 2213–2225.

Mackay, D. K., Sherstone, D. A., and Arnold, K. C. (1974). "Channel ice effects and surface water velocities from aerial photography of Mackenzie River break-up." *Hydrological aspects of northern pipeline development*. Report No. 74-12, Task Force on Northern Oil Development, Environmental-Social Program, Northern Pipelines, Saskatoon, Sask., Canada.

Marcotte, N. (1984). "Anchor ice in Lachine Rapids: Results of observations and analysis." *Proc. IAHR Symposium on Ice*, 1, 151–159.

Martin, N., Roy, A. G., and Bergeron, N. E. (1996). "Structure of turbulent flow in an ice-covered tidal environment." *Proc. 2nd International Symposium on Habitat Hydraulics,* INRS-Eau, Quebec, A435–A445.

Martin, S. (1981). "Frazil ice in rivers and oceans." *Annual Review of Fluid Mechanics*, 13, 379–397.

Martinson, C. (1980). "Sediment displacement in the Ottauquechee River—1975–1978." *CRREL Special Report 80-20*, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, N.H.

Marusenko, Y. I. (1956). "The action of ice on river banks." *Priroda*, 45(12), 91–93.

Matousek, V. (1984). "Types of ice and conditions for their formation." *Proc. IAHR Ice Symposium*, International Association for Hydraulic Research, Delft, the Netherlands, 1, 315–327.

Mercer, A. G., and Cooper, R. H. (1977). "River bed scour related to the growth of a major ice jam." *Proc. 3rd National Hydrotechnical Conference*, Canadian Society of Civil Engineers, Toronto, Ont., 291–308.

Meyer-Peter, E., and Muller, R. (1948). "Formulas for bedload transport." *Second Congress International Association for Hydraulic Research*, International Association for Hydraulic Research, Delft, the Netherlands, 39–64.

Michel, B. (1971). "Winter regime of rivers and lakes." *CRREL Monograph III-IBa*, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, N.H.

Michel, B. (1972). "Properties and processes of river and lake ice." *Proc. The Role of Ice and Snow in Hydrology*, Publication No. 107, International Association for Scientific Hydrology, Grenoble, France, 454–481.

Michel, B. (1978). *Ice mechanics*. University of Laval Press, Laval, Que., Canada.

Milburn, D., and Prowse, T. D. (1998). "The role of an ice cover on sediment transport and deposition in a northern delta." *Proc. IAHR Ice Symposium,* International Association for Hydraulic Research, Delft, the Netherlands, 189–196.

Morse, B., Burrell, B., St. Hilaire, A., Bergeron, N., Messier, D., and Quach, T. T. (1999). "River ice processes in tidal rivers: Research needs." *Canadian River Ice Workshop,* University of Manitoba, Winnipeg, 388–399.

Muste, M., Braileanu, F., and Ettema, R. (2000). "Flow and sediment-transport measurements in simulated ice-covered channel." *Water Resources Research*, 36(9), 2711–2720.

Neill, C. R. (1976). "Scour holes in a wandering gravel river." *Proc. Symposium on Inland Waterways for Navigation, Flood Control, and Water Diversions, 3rd Annual Conf. Of ASCE Waterways, Harbors and Coastal Engineering Division*, ASCE, New York, 1301–1317.

Neill, C. R. (1982). "Hydrologic and hydraulic studies for northern pipelines." *Proc. Canadian Society of Civil Engineering Annual Conference,* Canadian Society of Civil Engineers, Toronto, Ont., 247–256.

Newbury, R. W. (1982). *The Nelson River: A study of sub-Arctic processes*. John Hopkins University Press, Baltimore, Md.

Nezhikovskiy, R. A. (1964). "Coefficients of roughness of bottom surface of slush ice cover." *Transactions of the State Hydrologic Institute* (Trudy GGI), (110), 54–82.

Osterkamp, T. E., and Gosink, J. (1983). "Frazil ice formation and ice-cover development in interior Alaska streams." *Cold Regions Science and Technology*, 8, 43–56.

Prowse, T. D. (1993). "Suspended sediment concentration during river ice breakup." *Canadian Journal of Civil Engineering*, 20, 872–875.

Prowse, T. D. (2001). "Rice-ice ecology. I: Hydrologic, geomorphic and water-quality aspects." *ASCE Cold Regions Engineering*, 15(1), 17–33.

Raudkivi, A. J. (1998). *Loose boundary hydraulics.* A. A. Balkema, Rotterdam, the Netherlands.

Reid, J. R. (1985). "Bank-erosion processes in a cool-temperate environment, Orwell Lake, Minnesota." *Geological Society of America Bulletin*, 96(6), 781–792.

Reimnitz, E., and Kempema, E. W. (1987). Field observations of slush ice generated during freeze up in arctic coastal waters. *Marine Geology*, 77, 219–231.

Reynolds, A. J. (1974). *Turbulent flows in engineering*. Wiley, New York.

Sayre, W. W., and Song, G. B. (1979). "Effects of ice covers on alluvial channel flow and sediment transport processes." *IIHR Report No. 218*, Iowa Institute of Hydraulic Research, The University of Iowa, Iowa City, Ia.

Schumm, S. A. and Khan, H. R., (1972). "Experimental Study of Channel Patterns."*Geological Society of America Bulletin,* 83, 1755–1770.

Scott, K. M. (1978). "Effects of permafrost on stream channel behavior in Arctic Alaska." *Professional Paper No. 1068*, U.S. Geological Survey, Washington, D.C.

Scrimgeour, G. J., Prowse, T. D., Culp, J. M., and Chambers, P. A. (1994). "Ecological effects of river ice break-up: A review and perspective." *Freshwater Biology*, 32, 261–275.

Shen, H. T., and Wang, H. Y. (1995). "Under cover transport and accumulation of frazil granules." *Journal of Hydraulic Engineering, ASCE*, 121(2), 184–195.

Simon, A. Shields, F. D., Ettema, R., Alonso, C., Marshall-Garsjo, M., Curini, A., and Steffen, L. (1999). "Channel erosion on the Missouri River, Montana, between Fort Peck Dam and the North Dakota Border." *Technical Report*, USDA-Agricultural

Research Service, National Sedimentation Laboratory, Oxford, Miss.

Smith, B., and Ettema, R. (1997). "Flow resistance in ice-covered alluvial channels." *Journal of Hydraulic Engineering, ASCE*, 123(7), 592–599.

Smith, D. G. (1979). "Effects of channel enlargement by river ice processes on bankfull discharge in Alberta, Canada." *Water Resources Research*, 15(2), 469–475.

Straub, L. G. (1955). "Effect of water temperature on suspended sediment load in an alluvial river." *Proc. 6th General Meeting of IAHR*, International Association for Hydraulic Research, Delft, the Netherlands, D25-1–D25-5.

Sukhodolov, A., Thiele, M., Bungartz, H., and Engelhardt, C. (1999). "Turbulence structure in an ice-covered, sand-bed river." *Water Resources Research*, 35(3), 889–894.

Taylor, B. D., and Vanoni, V. (1972). "Temperature effects in low-transport flat-bed flows." *Journal of Hydraulics Division, ASCE*, 98(HY8), 1427–1445.

Terada, K., Hirayama, K., and Sasamoto, M. (1997). "Field measurements of anchor and frazil ice." *Proc. IAHR Symposium on Ice*, International Association for Hydraulic Research, Delft, the Netherlands, 2, 697–702.

Tietze, W. (1961). "Uber die Erosion von unter Eis Fliessendem Wasser." *Mainzer Geographische Studien*, 125–141.

Tsai, W.-F., and Ettema, R. (1994). "Ice cover influence on transverse bed slopes in a curved alluvial channel." *Journal of Hydraulic Research*, 32(4), 561–581.

Tsang, G. (1982). *Frazil and anchor ice: A monograph*. National Committee on Hydraulics of Ice Covered Rivers, National Research Council of Canada, Ottawa, Ont., Canada.

Tuthill, A., and Mamone, A. C. (1997). "Selection of confluence sites with ice problems for structural solutions." *CRREL Special Report 97-4*, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, N.H.

Tywonik, N., and Fowler, J. L. (1973). "Winter measurement of suspended sediments." *Proc. IAHS Conf. on the Role of Snow and Ice in Hydrology*, International Association for Hydraulic Science, Grenoble, Switzerland, 814–827.

U.S. Army Corps of Engineers (USACE). (1983). "Galena streambank protection." *Galena, Alaska Section 14 Reconnaissance Report*, U.S. Army Corps of Engineers, Alaska District, Anchorage, Ak.

Uunila, L. S. (1997). "Effects of river ice on bank morphology and riparian vegetation along the Peace River, Clayhurst to Fort Vermilion." *Proc. 9th Workshop on River Ice*, Natural Research Council of Canada, Ottawa, Ont., 315–334.

Wentworth, C. K. (1932). "The geologic work of ice jams in sub-Arctic river." *Contributions in geology and geography*, L. F., Thomas, ed., Washington University Studies, Science and Technology No. 7, Washington University, St Louis, Mo., 49–82.

Wigle, T. E. (1970). "Investigations into frazil, bottom ice, and surface ice formation in the Niagara River." *Proc. IAHR Symposium on Ice and Its Action*, International Association for Hydraulic Research, Delft, the Netherlands, Paper 2.8.

Wuebben, J. L. (1986). "A laboratory study of flow in an ice-covered sand bed channel." *Proc. 8th IAHR Symposium on Ice,* International Association for Hydraulic Research, Delft, the Netherlands, 1–8.

Wuebben, J. L. (1988a). "A preliminary study of scour under an ice jam." *Proc. 5th Workshop on Hydraulics of River Ice/Ice Jams*, National Research Council of Canada, Ottawa, Ont., 177–190.

Wuebben, J. L. (1988b). "Effects of an ice cover on flow in a movable bed channel." *Proc. 9th IAHR Symposium on Ice*, International Association for Hydraulic Research, Delft, the Netherlands, 137–146.

Wuebben, J. L. (1995). "Chapter 31: Ice effects on riprap." *River, Coastal and Shoreline Protection: Erosion Control Using Riprap and Armourstone*, C. R. Thorne, S. Abt, S. T. Barends, and K.W. Pilarczyk, eds., Wiley, New York, 513–530.

Wuebben, J. L., and Gagnon, J. J. (1995). "Ice jam flooding on the Missouri River near Williston, North Dakota." *CRREL Report 95-19*, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, NH.

Yalin, M. S., (1992). *River Mechanics.* Pergamon Press, Oxford, Britain.

Zabilansky, L. (1998). "Scour measurements under ice." *Proc. ASCE Conf. Water Resources Engineering*, ASCE, Reston, Va., 1, 151–156.

Zabilansky, L., Ettema, R., Wuebben, J. L., and Yankielun, N. E. (2002). "Survey of river-ice influences on channel bathymetry along the Fort Peck reach of the Missouri River, winter 1998–1999." *CRREL Report* (in press), U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, N.H.