ENVIRONMENTAL ASPECTS OF RIVER ICE

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FOREWORD

The Task Force on Environmental Aspects of River Ice was formed in 1988 under the auspices of the Subcommittee on Hydraulics of Ice Covered Rivers. Its membership is as follows:

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The formation, persistence, and breakup of the ice cover in Canadian rivers can have considerable impact on aquatic ecosystems but very little pertinent information is available in the literature and it is often scattered in journals of very different disciplines. A major objective of the Task Force has been to review and summarize the state of the art and to point out areas where research is needed. This objective has been successfully met thanks to the efforts and contributions of the members of the Task Force.

It is hoped this report will help disseminate knowledge on the environmental effects of river ice and, at the same time, increase awareness of the importance of the subject and stimulate future research. To this end, the Subcommittee's seventh Workshop on River Ice (Saskatoon, August, 1993) will focus on environmental aspects and provide a forum for current work from different disciplines to be presented and discussed. The Workshop Proceedings will be a useful companion document to the present one.

S. Beltaos
Chair, Subcommittee on Hydraulics of Ice Covered Rivers

July 1993

PREFACE

River environments which undergo periodic formation of ice will have a morphology, a chemistry, and an ecology adapted to the ice regime and to the periods of freeze-up and debacle.

Economic developments on river basins which impact the natural river-ice regime will perturb the natural physical and chemical norms in the channel which in turn may have repercussions on the aquatic ecosystem and water quality.

Successful application of policies of sustainable development depends on first understanding river aquatic processes and successfully modelling them so that the proposed changes in regime are part of the input variables or boundary conditions. A particularly critical time occurs during periods of ice formation or melt which greatly complicates the physical regime.

This scientific report on the environmental aspects of river ice is the necessary first step in evaluating and describing in an integrated way all the physical, chemical or biological processes at work in the river during the critical periods when ice dominates. Much more research is needed to obtain reliable information on relationships between key variables.

Economic development will progressively impact on northern rivers and give rise to more and more requirements to evaluate the sustainability of the projects. Accumulation of impacts will be an issue. Scientists and engineers will be asked to assess and describe reliably impacts of development on the ice regime and the concomitant effects on the aquatic ecology, river water quality, and morphology. Currently, this cannot be done in a quantitative way.

Scientists or engineers who must evaluate and/or assess the effects of developments on rivers, especially where winter conditions may be critical, should find this state-of-the-art report useful for identifying where further studies, research and investigations should be focused.

The editors and contributors to this report begin an integrated scientific response to address questions on the impact of ice on river physical, biological, and chemical regimes. Their contribution and dedication are gratefully acknowledged.

T. Milne Dick
Director, National Hydrology Research Centre

July 1993

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INTRODUCTION

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1.1 BACKGROUND

Freshwater ice seasonally covers most lake and river systems of the cold regions ranging from periodic skims of ice in the more southerly temperate regions to mean thicknesses in excess of 2 m on high-latitude lakes and rivers. The effect of such ice covers has been the focus of scientific research for most of this century. The focus, however, has been primarily on hydraulic effects and only minor attention has been paid to the broader environmental significance of river ice. As utilization of cold-regions river systems for activities such as water supply, recreation and power generation has expanded, the significance of ice covers to the broader physical, chemical and biological environmental systems has become increasingly recognized. Much of this recognition has been brought about through post-development environmental impact assessments of major water-development schemes. To avoid potential future negative impacts, to protect cold-regions environments, and to manage more effectively cold-regions development, there is a strong need for more information about the environmental importance of river ice. Unfortunately, scientific publications dealing with these topics are relatively meagre, disparate, and spread throughout the literature.

Recognizing the lack of comprehensive information on river ice environmental problems, the National Research Council of Canada's Subcommittee on Hydraulics of Ice Covered Rivers recommended in

1988 that a Task Force be formed to investigate the environmental aspects of river ice. The Task Force has assembled this state-of-the-art report in order to document current knowledge and develop recommendations for research.

1.2 TASK FORCE

In June 1988, the National Research Council of Canada (NRCC) Subcommittee identified an initial membership of the Task Force. This group of individuals first met in November 1988 at the Canada Centre for Inland Waters in Burlington, Ontario. Additional potential members were identified and appointed. The Task Force membership now comprises scientists and engineers from across North America. A list of members is provided in the foreword to this report written by S. Beltaos, Chair of the Subcommittee on Hydraulics of Ice Covers.

1.3 ORGANIZATION OF THE REPORT

This state-of-the-art report comprises contributions by Task Force members and is organized under three major chapters: physical, chemical and biological. The editors have re-arranged original material and inserted new text to enhance the degree of coverage, to maximize consistency among authors' presentations, and to better link the various sections. A subsequent publication "Proceedings of the Workshop on Environmental Aspects of River Ice" will be published following the release of this science report at the workshop of the same name.

1.4 ACKNOWLEDGEMENTS

The editors are indebted to all the authors for their excellent contributions to the report. Special thanks are also due to the various people who made it possible to complete the publication. In particular, much credit is due to Janice Akre, Karen Sollid, Jane Thul, and Leah Watson, all of the National Hydrology Research Institute (NHRI), for their assistance in management and production of the report. The editors also wish to acknowledge the efforts and talents of Philip Gregory (NHRI) who redrafted many of the figures and designed the very striking cover.

2

PHYSICAL EFFECTS OF RIVER ICE

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Editors' Note: In editing this chapter, it was necessary to rearrange original material and write new text to enhance the degree of coverage, to maximize consistency among authors' presentations, and to better link the various sections. Hence, it was very difficult, in many cases, to assign final authorship to specific sections. The above listing, however, indicates the name of all contributing authors and the sections for which they were major authors. Sole-authors and first-authors are denoted by a bolded section number. Other major contributing authors are shown by a standard-font section number. Even this method does not recognize authorship of some minor pieces, but it was decided to be the fairest way to indicate overall authorship.

2.1 INTRODUCTION

This chapter focuses on the physical role of river ice. As background for readers not familiar with the field, it first describes some major river ice types and covers, and how these are formed by and affect instream hydraulic processes. The discussion is then expanded to include some of the broader hydrologic effects, especially as they relate to seasonal and annual hydrographs. The influence of ice on the mixing and transport of substances is reviewed next, including sections on sediment transport and scour. Two subsequent sections elaborate on the erosional and depositional effects of ice in producing unique shoreline and substrate features and characteristics. A review of the effect of ice on river thermal characteristics, heat balances and local climate concludes the chapter. Areas requiring future research are noted throughout.

2.2 MAJOR RIVER ICE TYPES AND COVERS

2.2.1 Introduction

River-ice covers usually comprise a myriad of ice types, formed by a complex array of static and dynamic growth/accumulation processes. As background for more detailed subsequent discussions, the following sections review the major river ice and cover types, focusing primarily on their genesis and hydraulic effects. Some examples of the broader environmental significance of various ice types are given, but more detailed discussion of these comes in later sections and chapters.

2.2.2 Frazil Ice

The bulk (average) water temperature in a stream has to be close to the freezing point before significant ice can be produced. Surface-type ice will be the first to form along the shorelines even when the water temperature in the main flow may be above freezing. In the faster flowing sections, which are well mixed and turbulent, frazil ice can form throughout the water column.

Frazil ice is only generated in a turbulent flow when the water temperature is slightly depressed a few hundredths of a degree below the freezing point. During this growth period the frazil ice particle is particularly "sticky" and will attach to almost any object, including other ice crystals; this is often referred to as the active state. The effect of the growth of frazil ice (discoids) in the water column on the flow regime has not been easy to determine, simply because of the difficulty in measuring the thermal and hydraulic conditions during these periods. Chacho *et al.* (1986) video-recorded the transport of "non-active" or passive frazil ice under the Tanana River. The vertical distribution of frazil ice throughout the flow depth has been measured by Tsang (1985), Liu and Ferrick (1992), and Omstedt (1985) have formulated models for the vertical distribution of particles. It should be obvious

that the higher concentrations of ice particles will be near the surface because of buoyancy and flocculation of individual particles.

The flocculation and subsequent growth of the frazil particles on the surface that eventually transpose into composite solid/unconsolidated surface ice floes have been formulated recently by Shen and Wang (1992). The transport of frazil ice under the cover is related to the rise velocity of the particles and the carrying capacity of the flow to transport the suspended ice. Transport equations originally developed for sediment movement have been applied to ice transport under stationary ice covers.

There has been some speculation that organic material suspended in the water column enhances the nucleation of frazil ice, but in studies performed by Mueller and Calkins (1978) of supercooled water temperatures near -0.05°C, no evidence of this was found. It has also been observed by various investigators that turbid waters seem to clear during periods of frazil ice generation; the reasons for this are not clear, though one could speculate that during the frazil ice growth period the "turbid" material was scavenged by the frazil discs.

2.2.3 Anchor Ice

Anchor ice is defined as ice either growing or depositing on the substrate of the river bed. This ice may consist of frazil discoids or needle-like crystals (e.g., see Tsang, 1982). Both processes have been observed to occur independently of each other or they may occur simultaneously, as indicated by the crystalline structure of the deposited material. Although found adhering to almost any underwater object, anchor ice is most commonly found sticking to aquatic vegetation, boulders and even large areas of gravel and coarse sand. In small streams, such as found in northern New England of the United States, depending on the local thermal/ice hydraulic conditions, anchor ice can eventually become the major component of the solid surface ice cover.

The effect of anchor ice on the flow hydraulics is twofold. First, the deposition and growth of anchor ice on the substrate can create a sufficiently thick false bottom that the river stage can rise above its normal depth for the same discharge in the absence of the anchor ice. Second, intergravel flow that occurs between the surface flow and the substrate flow can become blocked. Anchor ice fills the voids at the substrate surface and effectively cuts off the flow paths for the exchange of surface and subsurface waters (Calkins, personal communication). This can have serious biological implications for biota sensitive to freezing temperatures or dependent on the oxygen supplied by intergravel flow (e.g., Blachut, 1988; Calkins, 1989; see Chapter 4). Field studies by fisheries researchers on the flow of water through spawning redds can be traced back as far as the 1950s (Stewart, 1953), while hydraulicians have been studying the flow over porous media boundaries for about the same period; it appears that the two disciplines are closing the gap. Wang and Shen (1991) have put forward a relationship that describes the growth and decay of anchor ice on a porous river substrate in terms of heat transfer relationships. Coupling the equations that describe the mass flow and heat fluxes through the substrate is the next logical step.

2.2.4 Static Ice

Once an ice cover becomes established (see discussion of cover types below), it begins to grow into the underlying water, typically forming a relatively-transparent columnar ice, similar to that occurring on lakes. The growth and appearance of this ice, however, can be modified by the presence of other dynamic ice forms such as frazil ice. Layers of snow serve to retard growth but can also accelerate it through a process of slushing, whereby it is incorporated into the ice sheet as snow-ice or white-ice. Such surface forms of ice increase the overall ice thickness but also decrease the downward growth of static ice growth because of their added insulation factor. For reasonably thick ice sheets, a depth of snow approximately equal to half the ice-sheet thickness will halt the growth of static ice (Michel, 1971). Notably, however, as found by Adams and Prowse (1981), in lakes, the decrease in black ice growth due to insulation created by a snowcover is usually compensated for in the long term by the additional ice thickness created by snow-ice. Although no comparable studies of spatial variability in ice stratigraphy have been conducted on rivers, a similar relationship is likely to exist. As discussed in Section 2.8, the composition of the various ice forms can significantly affect the heat budget and underwater spectral regime.

2.2.5 Icings

In addition to anchor ice (Section 2.2.3), icings are another type of ice accumulation that can strongly affect the channel bed of streams and rivers (*aufeis* is the German term, *naleds* the Russian term — Sokolov,

1978). They can form on existing river-ice covers or on the land surface where groundwater discharge emerges and freezes (Carey, 1970 and 1973; Kane and Slaughter, 1972; Kane, 1981). Icings can represent the total winter discharge of the base flow (groundwater) component of a basin, as evidenced on the north slope of the Yukon, Canada, where the Babbage and Firth rivers have no measurable flow for five to six months each year (van Everdingen, 1974). The calculated flow rate from groundwater for these streams is 1.4 m^3/s, which gives an indication of the volume of water going into seasonal storage in the form of icing deposits.

Within main flow channels, icings can also develop in cases where progressive vertical growth of ice creates enough of a flow obstruction to force water to the surface, eventually to freeze and create surface ice-forms. Depending on the surface flow rate and the rate of heat loss, the exposed water will refreeze as a large ice mound or hummock near the point of seepage, or as a large sheet of ice, spreading some distance downstream. Water may also emerge from a number of locations to form one large icing that may extend across the entire channel width and even spread out onto the surrounding floodplain. On some streams, this redirection of flow can effectively remove the baseflow component of runoff from the channel, thereby reducing the instream flow and accentuating the bed freezing-process.

Where the icing process continues throughout the winter, many arctic and subarctic streams develop ice thicknesses several times the normal depth of open channel flow recorded during the summer months. Moreover, the thickness of some arctic icings is two to three times the maximum thickness of thermal ice covers formed on deep rivers and lakes. Shumskii (1964), for example, reported some Siberian icings being 5-10 m thick, 27 km long and containing up to 5.0×10^8 m^3 of ice. Satellite imagery, such as from the early Landsat (Land Remote-Sensing Satellite; e.g., van Everdingen, 1976) and the more recent Landsat-TM (thematic mapper), has been used to map large Arctic icings. The new Synthetic Aperture Radar images from ERS-1 (European Remote-Sensing Satellite), and the soon-to-be launched Canadian Radarsat, will expand this capability.

During periods of significant run-off, icings largely determine the channel routing and can act as major flow restrictions. They are of special concern where flow is routed through narrow channels or culverts. Thick icings occupying the full channel width also increase the flood potential by reducing the flow required to reach bankfull stage and possibly diverting the spring flow out of the normal channel margins. The protection they offer the channel bed, however, can also reduce the effect of the erosive processes that normally occur at this time (Kane and Slaughter, 1972; also see Sections 2.6 and 2.7).

Large icings can persist long after the anticipated break-up of a floating ice cover and, through their subsequent melt, lead to a significant seasonal redistribution of flow. The previously cited icings on the Babbage and Firth rivers, for example, develop over six months and take approximately three months to melt, thereby increasing late-spring and summer streamflow by 1.0 to 2.8 m^3/s (van Everdingen, 1987). This type of protracted melt has special environmental significance for rivers in northern climates where summer flows are normally low because of the lack of precipitation. In some cases, major icings survive for several years before fully melting (Grey and MacKay, 1979).

In addition to providing a source of extended flow, such icings often have a unique geochemistry representative of the subsurface strata from which they originate (van Everdingen, 1990). Where these icings provide the main flow during dry summer periods, they also strongly control the stream water quality (van Everdingen, 1987). They may also act as a natural source of minerals for wildlife such as caribou (e.g., Edwards and Ritcey, 1960; Skoog, 1968).

2.2.6 Surface Ice Cover - Moving

While anchor ice and icings are special forms of river ice "covers", particularly on small streams, the primary ice cover on most larger streams and rivers develops on the surface of the water column from the growth and accumulation of various static (Section 2.2.4) and dynamic (Section 2.2.2) ice types. During the initial freeze-up phase, surface ice can develop as a moving cover. The effects of such a cover on the flow regime are complicated, however, because of the potential feedback process between a moving mass of ice and the moving flow.

When ice is being transported downstream and is moving relatively unimpeded without many major surface obstructions, the drag exerted by the moving ice to a steady flow is small. If the streamflow is accelerating or decelerating (unsteady flow), the resistance to the flow will be increased somewhat. The flow of water and ice at the surface will be somewhat less than the maximum velocity of the fluid, which occurs some distance below the surface.

The movement of floating ice on the water surface is influenced by several factors. The wind drag on the surface ice in wide rivers with sufficient fetch can drive the floes to one side of the channel, secondary flow currents that are induced in river bends will move the surface ice toward the outside, and emerging flows from tributaries can change the flow pattern in the main stem river and alter the surface ice pathlines. Withdrawals of water from streams can also affect the direction of a moving ice cover. The hydraulics of some of these occurrences have been analyzed in the past and some are only now under study using particle trajectory analysis that includes floe collision; it is not a simple problem of analysis, as the factor of unsteady flow is apparent in many of these conditions.

The hydraulic situations in a natural river system that arise when moving ice comes in contact with a stationary cover will depend on the quantity of incoming ice and the reaction of the stationary ice cover to this incoming ice: i.e., does the initially stationary cover remain stable or become mobilized? River waves will result from this abrupt change when a stationary cover makes the transition to a moving cover. Depending on the situation, upstream, downstream, or both types of propagating waves will develop. Their magnitude and duration are a function of many hydraulic and ice properties; see Yapa and Shen (1986) or Ferrick *et al.* (1992) for more complete descriptions of water waves in streams and rivers.

An ice sheet normally at rest but suddenly mobilized initiates a major hydraulic change in the river at the location where the mobilization occurs, and this change in velocity will be propagated both upstream and downstream. The change in the upstream direction is a negative wave (depths will be decreasing) while in the downstream direction a positive wave of water and ice will develop, with depths increasing. The velocity of the ice and water in the positive wave can easily reach 4-5 m/s. If the surge of water and ice is then halted, another upstream wave can be expected, this being a positive wave (increasing depths) where the change can be readily seen by an observer. Rapid rises of several metres within minutes is common on some of the larger rivers. The impacts of these high-velocity conditions on the riverbed and riverbank are bank erosion and substrate degradation with eventual sediment deposition at some downstream location (see Sections 2.5 to 2.7 regarding scour and effects on shorelines and substrates).

2.2.7 Surface Ice Covers - Suspended

For many small headwater streams, the computation of classical open channel hydraulics with floating ice covers will not provide an accurate description of the flow regime; it may not even be close because of the highly variable ice thickness across the channel. Because of the dynamic nature of the meteorological and ice conditions, the flow regime under the ice cover will be subject to change. In cobble-strewn streams, the streamflow in contact with both the bed and ice-cover underside during the freeze-up period can change to a flow situation where it is in contact only with the bed and an air gap is present under the ice cover. This can occur for a couple of reasons: the ice cover becomes supported by the cobbles and boulders and when the winter flow decreases, or when the warmer groundwater melts the underside of the ice cover, the stage drops and an air void develops between the bottom of the ice and the water surface (Calkins and Brockett, 1988). This air gap can have a tremendous insulating effect. The micro-climate of such cavities can form a unique habitat, especially when warmed by solar radiation penetrating the overlying ice sheet. On larger rivers, winter flow-recession is characterized by a concomitant lowering of ice and water levels except at shore contact-points. The space under hanging shore ice also forms a unique micro-climate and habitat. Muskrat, for example, are known to use such sub-ice spaces to forage channel-bank vegetation (Geddes, 1980). Other mammals (e.g., mink, otter and beaver) have also been observed in these zones between the ice and water.

2.2.8 Surface Ice Cover - Pressurized Situations

Pressure flow in streams is a rare occurrence because the ice cover is not a perfect containment surface. The ice cover has cracks, and it flexes, floats or even breaks when the flow is increased beneath it. However, there are times when the ice cover can experience pressure flow. During periods of intense run-off, such as from rain-on-snow events, rapid snowmelt and/or spring flows, streams with ice covers rigidly attached to the banks can experience a short-duration pressure flow. This pressure flow may not last a very long time, due to cracks in and creeping of the ice, and may only produce excess pressures of less than 3 KPa, which is probably not significant.

Another situation in which water in rivers can be under pressure occurs in the streams of the far north. Arcone *et al.* (1989) and Chacho (1990), for example, have documented river ice mounds on several streams in the Arctic National Wildlife Refuge in northern Alaska. These mounds occur in streams across the entire circumpolar region. Some mounds were still under pressure when drilled, and in some cases the hydrostatic pressure was roughly 25 KPa. Several mounds were observed to have cracked and partially drained, as evidenced by the overflow icing on the side of the mound. As reported in Arcone *et al.* (1989), Soviet studies have reported that native villagers have heard these mounds rupture. Continuing studies by researchers at the U.S. Army Cold Regions Research and Engineering Laboratory (USACRREL), of rivers where the ice cover is completely frozen to the bed, have recently documented the appearance of flow channels 3-4 m beneath the frozen substrate using ground penetrating radar. The biological activities or implications in these pressure situations and in these deep flow channels are unknown, although larvae have been observed in the cavities of the river mounds after they have drained.

2.2.9 Surface Ice Cover - Stationary

In rivers that are relatively wide (>100 m) with ice thickness generally less than 50% of the total river depth, the effect of a stationary floating ice cover on the flow hydraulics can be computed with confidence and has been well documented from field data (e.g., Ashton, 1986).

The basic effect of a floating stationary ice cover is that the underside of the ice offers additional resistance to the flow, thereby increasing the depth required to pass that flow. The magnitude of the increase in flow depth (measured from the bed to the ice cover underside) is a function of the roughness of the underside and the change in the wetted perimeter of the flow due to the additional ice surface. If the roughness of the bed and ice is similar, then an increase of roughly 30% can be expected. The following sections elaborate on the hydraulic relationship between ice and water levels, and the effects of backwater conditions.

2.3 HYDRAULIC EFFECTS OF AN ICE COVER

2.3.1 Introduction

The flow of water in a river ice environment can be analyzed using concepts developed for open channel flow hydraulics. The flow conditions of steady or unsteady, uniform or non-uniform can all occur in the presence of floating and stationary ice covers. The description of the hydraulic characteristics with an ice cover given below assumes that the flow is steady and uniform (constant flow and section properties). Although these are ideal conditions, they do provide the framework for a basic understanding of the interaction of the flow regime with an ice cover. The intent of this section is not to review the theoretical deviations for flow with ice covers, but to identify the major roles that an ice cover plays in affecting the hydraulics of streams and rivers, and to highlight some of the related, broader environmental impacts. Discussion is also included about current ice-affected modelling techniques.

2.3.2 Hydraulic Effect on Water Levels

The most commonly used formulation for the description of the flow conditions in a stream is the Manning equation, although other resistance equations such as the Chezy and the Darcy may also be used. The Manning equation is presented because many of the instream habitat models presently use it as their equation to describe the flow depth or velocity characteristics.

The Manning equation in SI form is:

$$V = \frac{1.0}{n} * R^{0.67} * S^{0.5} \qquad (2.1)$$

where V = flow velocity, R = A/P = hydraulic radius, A = cross sectional area, P = wetted perimeter of the flow, S = water surface slope, and n = Manning roughness coefficient. The term 1/n is assumed to have units of $m^{-1/3}/s$.

By rearranging the Manning equation and assuming the river top width is approximately equal to the wetter perimeter of the bed, the uniform flow depth (Y_0) in a wide rectangular channel with no ice is:

$$Y_0 = \left[\frac{Q n_b}{B \sqrt{S_0}} \right]^{3/5} \qquad (2.2)$$

where Q = flow discharge, n_b = bed roughness coefficient (Manning), B = top width of river, and S_0 = uniform flow slope.

The presence of a floating ice cover requires that equation [2.2] be modified to account for (a) the increase in the channel's wetted perimeter from the ice cover underside, (b) the area of the channel cross section occupied by the floating ice cover, and (c) the additional roughness of the underside of the ice cover. Figure 2.1 is an idealized view of this situation.

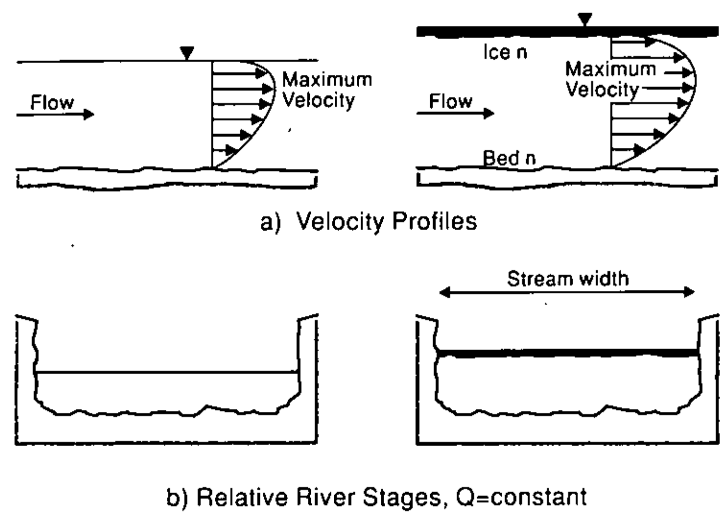


Figure 2.1 River stage - open and ice covered situations

The hydraulic radius, R, for an ice-covered wide rectangular channel may be approximated by:

$$R = \frac{A}{P+B} = \frac{A}{2B} = \frac{Y_i}{2} \qquad (2.3)$$

where P=B, and Y_i is the mean depth measured from the bed to the ice cover underside.

To account for the additional roughness of the ice underside, a composite Manning roughness coefficient, n_c, can be computed from the simple relationship:

$$n_c = \left[\frac{n_i^{3/2} + n_b^{3/2}}{2}\right]^{2/3} \qquad (2.4)$$

where n_i is the hydraulic roughness coefficient of the ice cover.

Typical values for n_i for a single-layer ice sheet range from 0.010 (e.g., mid-winter smooth cover) to 0.035 (e.g., pre-break-up thermally-roughened cover). Values for a rough ice jam surface may approach 0.1. References to additional typical values can be found in Ashton (1986) and Gray and Prowse (1993).

Making the substitution for the change in the hydraulic radius, equation [2.3] is substituted into equation [2.1] and the normal depth of flow in a wide rectangular channel beneath an ice cover with uniform flow (Y_i) becomes:

$$Y_i = 1.32 \left[\frac{Qn_c}{B\sqrt{S_0}}\right]^{3/5} \qquad (2.5)$$

where the factor 1.32 only accounts for the change in the hydraulic radius, it being approximately $Y_i/2$. An interesting result of the above equation is that the depth of flow beneath an ice cover, for uniform flow, is independent of the ice-cover thickness.

Hence, for a wide rectangular channel the total water depth, H, is:

$$H = 1.32 \left[\frac{Qn_c}{B\sqrt{S_0}}\right]^{3/5} + \varrho_i\, t_i / \rho_w \qquad (2.6)$$

where ϱ_i and ϱ_w are the densities of ice and water respectively, and t_i is the ice thickness.

If we assume that the bed and ice-cover underside have identical roughnesses, then equation [2.5] predicts that the flow depth will increase 32% from its open water depth with no ice at the same stream discharge. The ice cover could be 2 cm or 2 m, but the flow depth Y_i will remain the same, providing the roughnesses are the same. The total stage in an ice covered stream is composed of the flow depth plus the submerged thickness of the ice cover as given in equation [2.6].

Models to calculate the above hydraulic conditions are reviewed next followed by assessments of some of the backwater effects of ice-enhanced water levels.

2.3.3 Numerical Models: Ice Hydraulic Simulations

There are several computer programs available for computing water surface elevations due to the presence of a stationary, floating ice cover. HEC-2 from the U.S. Army Corps of Engineers (U.S. Army Corps of Engineers, 1979) is one of the most popular for analyzing gradually varied flow situations in river environments with ice covers, although it cannot analyze unsteady flow situations. For more background information about unsteady flow/ice conditions, readers are referred to general overviews in Ashton (1986) and specific studies including, for example, Beltaos and Krishnappan (1982), Ferrick *et al.* (1992), and Henderson and Gerard (1981). One- and two-dimensional ice/hydraulic models with unsteady flow capability are described in, for example, Yapa and Shen (1986) and Lal and Shen (1991). Comprehensive numerical models for ice/hydraulic simulations are currently being developed in Canada (Petryk *et al.*, 1991; Beltaos 1993a) and Finland (Huokuna, 1990).

With reference to hydraulic modelling of river ice, it is useful to put the previous discussions of "idealized" river ice/hydraulic situations into a realistic perspective. River ice covers are rarely uniform in nature, although conditions arise where the ice cover thickness can be relatively consistent in mildly-sloped rivers with low-flow velocity. The stationary ice cover in a river system is not a static parameter; it can consist of solid-ice and unconsolidated-ice fragments within its vertical dimension. The ice-cover thickness can also vary laterally across the stream as well as in the longitudinal direction. Depending on the hydraulic and thermal regimes of the river and the atmospheric conditions, the ice cover thickness will change throughout the winter season, potentially thinning in some areas and thickening in others.

The thickness of the ice cover in rivers can be a difficult value to predict unless one has a tremendous amount of hydraulic and meteorological data for the river of interest; to date the best one can do is a one-dimensional longitudinal profile. Although a two-dimensional simulation model is certainly possible to develop, the field data to verify such a model is time-consuming to collect. It has been documented in many studies that measurements of the ice regime over one to two seasons with relatively "normal" flow and meteorological conditions will provide enough information to show the temporal and spatial ice/hydraulic processes in the stream of study and will guide the necessary detailed studies if required at later dates.

The interaction of the ice cover and the flow hydraulics is a constantly evolving phenomena. During the freeze-up and break-up period, the dynamics of the interaction can be clearly visible; the cover can often be seen collapsing or thickening, water level changes are often rapid with over-bank flow common, and the "unsteady" situation of flow and ice movement is very evident.

It is fair to state that the measurement, evaluation, calculation, or simulation of the hydraulics of ice covers on larger rivers (widths greater than 50-100 m) have been studied extensively and there is a wealth of information from various studies on such rivers. The present knowledge gap about river-ice covers lies in the small rivers and streams, with flows that are difficult to characterize because of their highly-variable geometry, and in rivers that experience an ice cover of 7-8 months. The thermal regimes of the groundwater and the substrate in such rivers will play an extremely important role.

2.3.4 Backwater Effects

The ice-induced backwater effects described in Section 2.3.2 cannot only impact the entire winter flow-regime but can also have numerous important implications for the stream ecosystem, including overwintering fish and other aquatic animals, sediment transport, and riparian vegetation.

Some initial backwater can be created by, for example, the formation of ice on the streambed (anchor ice), and/or at the stream margins (shore ice) and/or in the water column (frazil ice). In small width (<20-30 m), steep (S_o > 0.01) streams, anchor-ice formations quite often extend to the water surface and water depths can easily double their open-water values. The duration of these high-stage increases may last only a few days and then subside due to flow recession and thermal melting or possibly last several weeks if the flow is stable and suface heat losses continue to be large. In general, water levels tend to increase further when a full stationary cover is in place. On the Nechako River (Blachut, 1988), for instance, backwater levels can be classified into two distinct regimes, depending on the type of ice: a) a 30 to 40-cm stage rise when anchor, frazil or shore ice was present, and b) an approximately 1-m stage rise when a full ice cover was in place (Figure 2.2).

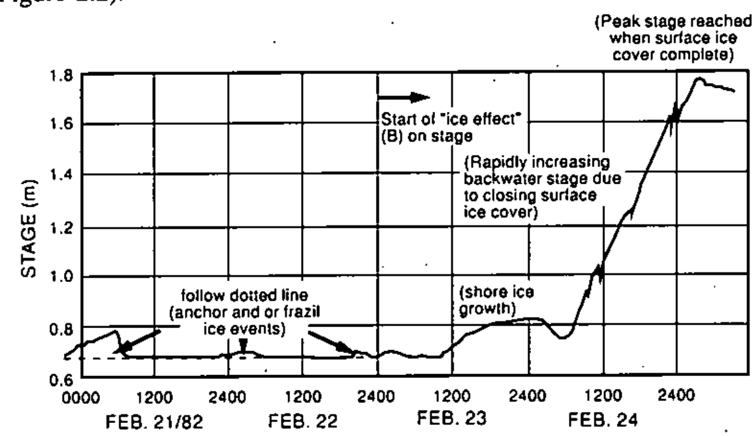


Figure 2.2 Trace of Water Survey of Canada stage recorder chart for Nechako below Cheslatta Falls Station (Water Survey of Canada, unpublished data)

The general effect on an open-water rating curve (stage-discharge relationship) is to raise the level component as exemplified in Figure 2.3; the calculation of which for operational applications has been the focus of considerable effort (e.g., Chin, 1966; Lavender, 1984; Rantz *et al.*, 1982; Rosenberg and Pentland, 1966). The magnitude of a backwater effect can be highly variable both spatially and temporally, varying primarily with the type and roughness of the ice. As discussed in Section 2.2.9 and 2.3.2, even when a complete cover is established, the backwater effect can continue to vary throughout the winter, primarily because of changes in bottom ice roughness. It tends to be high during the initial stages of freeze-up, decreases during the main period because of various smoothing processes, and then rises again during the pre-break-up period because of thermally-induced rippling.

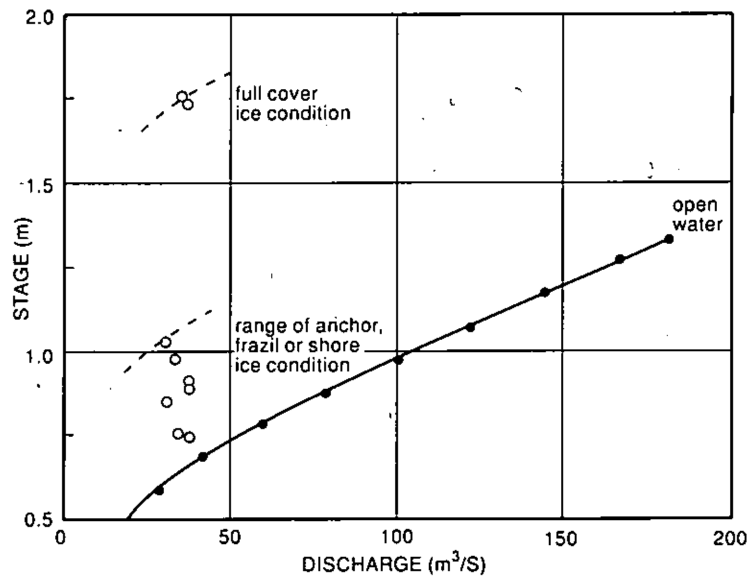


Figure 2.3 Nechako River below Cheslatta Falls (1985), open water and ice rating curves (Blachut, 1988)

Some of the most noticeable effects of freeze-up backwater occur in the near-shore and floodplain zones. As the flow depth and wetted perimeter increase, the previously exposed near-shore margins become inundated with backwater. Once an ice layer develops here, it can act as an insulator to the channel margin substrate and any biota present.

The insulation effect can be further enhanced with the addition of surface snow. However, if the backwater inundation area, which is typically quite shallow, freezes completely, then frost penetration into the substrate may be increased. This topic has rarely been investigated and its impact on overwintering conditions is unknown.

Increased stage may also inundate secondary channels not connected to the main channel prior to freeze-up. These include secondary channels, sloughs, or abandoned floodplain channels that can be connected hydraulically to the main channel by subsurface flow or that are only inundated when some threshold water level is exceeded (Paschke and Coleman, 1986). The thermal regime of off-channel areas is often very different from that of the main channel. Warm groundwater seeps are often present, mitigating the influence of colder mainstream flow. Warmer water has several effects: the thermal reserve keeps ice cover thin, prevents bed freezing, and provides better conditions for overwintering biota.

To satisfy the rise in water levels associated with ice-induced backwater, water is removed from the flow regime and placed into storage. This can have pronounced impacts on the seasonal hydrograph as discussed in Section 2.4.2.

2.3.5 Effects of Flow Diversion

In addition to the flow redistribution created by backwater effects described in Section 2.3.4, the simple presence of ice can cause other significant flow diversions. Anchor ice, for example, has been documented as the cause of significant flow diversions, albeit of relatively short duration (Figure 2.2). This diversion from selected areas of the streambed can have implications for biota overwintering in the substrate. Substantial amounts of sediment movement and geomorphic change have been observed to occur in an anchor ice event of less than one week's duration (Wilkins, unpub. data).

In the case of floating ice covers, the most common form of flow diversion develops when ice occupies significant portions of the channel cross-section. In zones of high heat loss and/or shallow flow, static-ice growth can totally occupy the channel depth, forming bottomfast ice. Estuarine locations and portions of braided channels are particularly susceptible (Anderson and MacKay, 1973). Formation of such ice along the shore zone can be significant, depending on channel shape, and results in a flow concentration to a smaller channel cross-sectional area. In extreme climates, such as the Mackenzie Delta (Northwest Territories, Canada), this flow confinement has resulted in a channel-within-a-channel morphology (Lapointe, 1985). The Peace-Athabasca Delta (Alberta, Canada) is another example, with channels that are seasonally blocked by ice, since complicated by hydroelectric regulation (Farley and Cheng, 1986; Peace-Athabasca Delta Project Group, 1973). Where ice freezes to the bottom of lake exit-channels, it can even block winter outflow (e.g., Peterson *et al.*, 1981). Given the sensitivity of lake biota to winter water-level fluctuations, such blockage has important environmental implications.

In its most extreme form, complete channel blockage results in free water being forced to the ice surface, resulting in the formation of overflow ice or aufeis (Forbes, 1979; Kane, 1981) as previously described in Section 2.2.5. The presence of all forms of bottomfast ice (e.g., static ice, anchor ice and icings) can also create unique flow situations during melt periods and a number of hydraulic and hydrometric-measuring complications (e.g., Egginton, 1978; Forbes, 1979).

Exceptionally thick ice growth can also be created by the additions of frazil ice deposited beneath a static ice cover (e.g., Calkins, 1979). Hanging dams are the extreme example of this process. The most common locations for their development include river mouths entering lakes or reservoirs, and deep pools within rivers, especially where they adjoin an upstream stretch of rapids in which large quantities of frazil are produced. Conditions where up to 40% of the cross-section is blocked by frazil accumulations are not uncommon (e.g., Alford and Carmack, 1987; Ashton, 1986; Beltaos and Dean, 1981). One of the most extreme examples is that found on the La Grande river in northern Quebec, Canada where one hanging dam was reported to be over 16 km long, comprised of over 56,000,000 cubic metres of frazil, and extending almost to the river bottom (Tsang, 1982). Such large frazil accumulations can not only obstruct and concentrate flow, but also lead to significant bottom scour. The formation of large holes within otherwise relatively uniform alluvial beds has been linked to the scour effects of hanging dams. During periods of increased discharge, the flow obstruction created by hanging dams can be severe enough to pose a flood hazard (e.g., Williams, 1971), as is discussed in the next section.

2.3.6 Ice-Related Flooding

As earlier reviewed, the obstruction and resistance to flow by river ice can lead to a diversion of flow and the inundation of adjacent portions of the channel. Ice can, however, also lead to widespread bankfull flooding, especially when the process of ice jamming is involved, either during freeze-up or break-up. Examples of ice-induced flooding abound in virtually all regions of Canada and the northern United States (e.g., Petryk, 1990) and their economic impacts are reasonably well documented (e.g., van der Vinne *et al.*, 1991).

While ice-related floods can be very destructive to man-made features, they can, on the other hand, be beneficial or even essential to riparian or deltaic environments. Key examples of the latter include the Mackenzie Delta (e.g., Blachut *et al.* 1985; Hirst *et al.*, 1987; Marsh, 1986; Marsh and Hey, 1989) and the Peace-Athabasca Delta (e.g., Peace-Athabasca Delta Project Group, 1973; Peterson *et al.*, 1981; Townsend, 1974; Wickware and Howarth, 1981; Prowse *et al.*, 1993). The ecosystem of such low-relief deltas has been found to depend on the spring flood produced by ice jams to recharge ponds and lakes. The warm, sediment-laden water that accompanies break-up, replenishes sediment and nutrients in the lake systems, and often forms the major seasonal input of the water balance for many of these lakes (e.g., Marsh, 1986). As discussed further below, this type of flood event is essential to the maintenance of the delta ecosystem, which can be considered a "pulse stabilized or fluctuating water level ecosystem" (Odum, 1971).

The mechanisms of delta lake flooding are relatively well understood for the Mackenzie Delta (e.g., Marsh and Hey, 1989) but to a far lesser extent for the Peace-Athabasca Delta (Peace-Athabasca Delta Ecosystem Management Plan, 1993; Peace-Athabasca Delta Implementation Committee, 1987; Peace-Athabasca Delta Project Group, 1973; Peterson *et al.*, 1981; Townsend, 1974; Wickware and Howarth, 1981). Other arctic or subarctic deltas with numerous lakes that would likely be subjected to similar ice-induced recharge include the Yukon, Colville and Blow rivers, Alaska (Dupre and Thompson, 1979; Walker and McCloy, 1969), Anderson River, Northwest Territories (Mackay, 1958), Babbage River, Yukon (Forbes, 1981), Slave River delta (Peterson *et al.*, 1981), Kvikkjokk and Laiture deltas, Sweden (Dahlskog, 1966; Axelsson, 1967) and the Yenisey, Lena, Ob, Kolyma and Indigirka rivers, Siberia (Antonov, 1969; Burdykina, 1970).

As is the case with flooding of ice-covered rivers, maximum water levels on delta lakes are usually dependent on ice-jam flooding that spreads from adjacent ice-covered channels and not simply on high-flow events. Lewis (1988) and Hirst *et al.* (1987), for example, found a poor relationship between annual-maximum river discharge of the Mackenzie River and annual-maximum water level in the adjacent Mackenzie Delta lakes. A series of studies conducted in this delta by the National Hydrology Research Institute (Bigras, 1985, 1990; Marsh, 1986; Marsh and Hey, 1988, 1989) on the hydrologic regime of delta lakes found that lake recharge was controlled primarily by lake sill-elevation and water levels (not discharge) in distributary channels. Furthermore, data from British Columbia Hydro studies in the same area (Blachut *et al.*, 1985) confirmed that the peak water levels occurred in conjunction with ice conditions, with peak stages ranging from one week prior to one week following clearance of ice from the immediate area (Figure 2.4). A correlation between ice thickness and spring flood elevations and timing has been demonstrated, where ice thicknesses greater than normal resulted in a 10-day delay in Mackenzie Delta break-up (Anderson and MacKay, 1973). The relationship between presence and proximity of ice and the peak water level is extremely dynamic, with considerable spatial and temporal variability. Certain types of lakes positioned higher on the floodplain are only flooded under extreme conditions. Observations of the 1982 break-up, an extreme event with an estimated return period of greater than one in 20 years, revealed that ice-jams resident for several days in the main distributary channels caused extensive flooding of the high floodplain (Blachut *et al.*, 1985). Thus, the presence of bottomfast ice in distributary delta channels (see Section 2.3.5) can also play a significant role in diverting flow and delta flooding (Anderson and MacKay, 1973).

Aspects of the ecology of ice-affected alluvial habitats have been studied in considerable detail, and relationships between various physical parameters and components of the ecosystem have been explored. Peterson *et al.* (1981) reviewed the ecological components of three northern Canadian deltas influenced by their ice regime: the Mackenzie, Slave, and Peace-Athabasca deltas. Vegetation communities and successional status as they relate to physical processes (ice-induced flooding) have been studied since the early 1970s in the Mackenzie Delta (e.g., Gill, 1971; Pearce, 1986; Hirst *et al.*, 1987; Wilkins and Hirst, 1989). Similar studies have been conducted also in the Peace-Athabasca delta (e.g., Peterson *et al.*, 1981; Peace-Athabasca Delta Implementation Committee, 1987). Maintenance of primary successional vegetation species such as willow by annual break-up

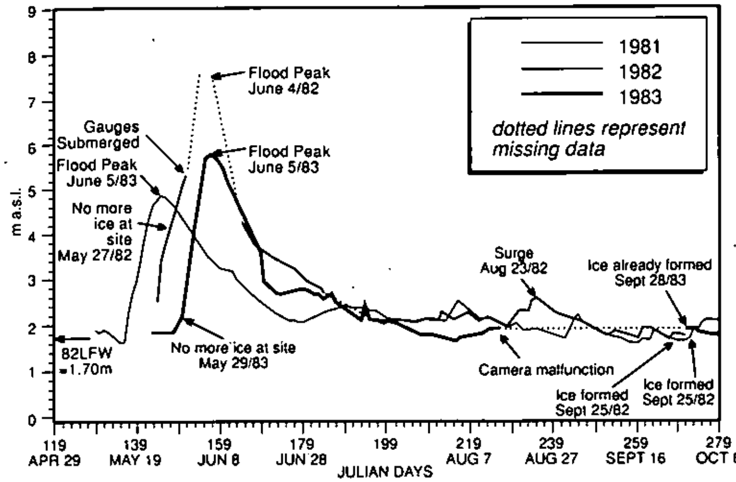


Figure 2.4 Water levels during and following ice break-up, Mackenzie Delta, 1981-1983 (Blachut et al., 1985)

flooding appears to provide good wildlife habitat. The delta lakes also are highly productive due to the influx of sediment and nutrient-rich floodwaters, resulting in dense populations of aquatic vegetation, and aquatic species such as fish, waterfowl and muskrats (see papers in Marsh and Ommanney, 1989). The presence or absence of ice has been suspected also of playing a role in the microclimate of such habitats (e.g., Gill, 1971; Hirst *et al.*, 1987).

2.4 WINTER HYDROLOGY

2.4.1 Introduction

The various water level disruptions created by ice-storage and ice-jam effects described in the previous sections can produce major features of the annual hydrograph including both floods and low-flow events (see hydrologic overview by Gerard, 1990). Seasonal redistribution of flows due to ice conditions can occur throughout the winter period and can be subdivided into three basic periods: freeze-up, mid-winter and break-up (Wedel, 1990).

2.4.2 Winter Hydrographs

One of the most pronounced effects of ice on winter hydrographs relates to the formation and release of water in backwater storage, as described in Sections 2.3.2 and 2.3.4. The storage effect is particularly noticeable during the freeze-up period when seasonal runoff is normally on a decline. The effect can be seen in water-level hydrographs by a rise in stage upstream of the ice-accumulation zone due to storage and a resultant decrease in downstream stage. During the freeze-up period the rise and fall of the stage can occur several times before a complete ice cover is reached. This effect is particularly noticeable on smaller streams in temperate regions (e.g., Santeford and Alger, 1986) but also occurs on larger, northern rivers (e.g., Gerard, 1990; Gray and Prowse, 1993). Regulated flows can make this process more difficult to observe.

The effects of storage are not only reflected in water level changes but also in flow, particularly at downstream locations (Figure 2.5). Typically, downstream discharge declines as flow is abstracted to satisfy the upstream storage requirement created by the accumulating ice cover. A portion of the abstraction can also be ascribed to ice growth but this is usually a small component except in arid areas where growth effects cannot be ignored. Where frazil ice growth is very strong, flow may even be slowed because of increased fluid viscosity created by a high concentration of ice particles (e.g., Tsang, 1979; Curtis, 1988).

Storage-related decreases in flow can occur quite abruptly, as when a moving cover bridges and begins to accumulate. The largest declines occur where ice storage is produced by a thick, rough cover such as a freeze-up jam, the effects of which are shown in Figure 2.6. The most pronounced effects are experienced on systems where lakes are present or where groundwater infiltration into alluvial streambank materials is possible (Gerard, 1989). So extreme are some of the freeze-up reductions in flow that they create a period of low flow, lower than that which is normally experienced in late winter when landscape runoff is at a minimum (e.g., Figure 2.5). Further discussion regarding low flows under ice are provided in Section 2.4.3.

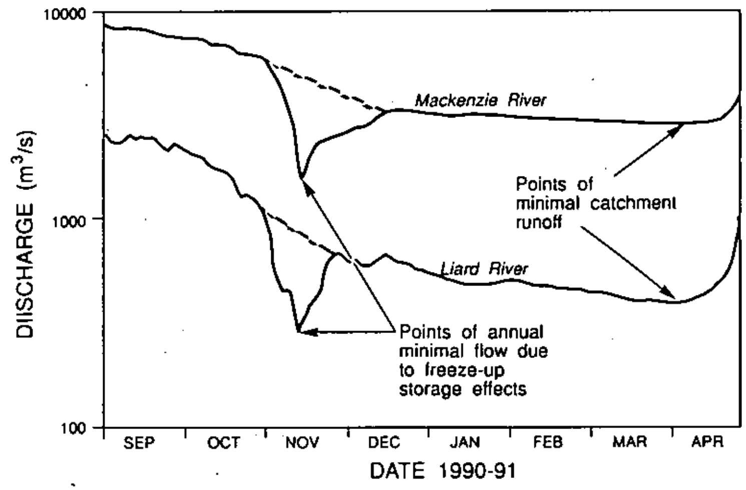


Figure 2.5 Winter hydrographs showing displacement of minimum flow due to freeze-up storage effects (data source: 1990-1991 Water Survey of Canada: from Gray and Prowse, 1993)

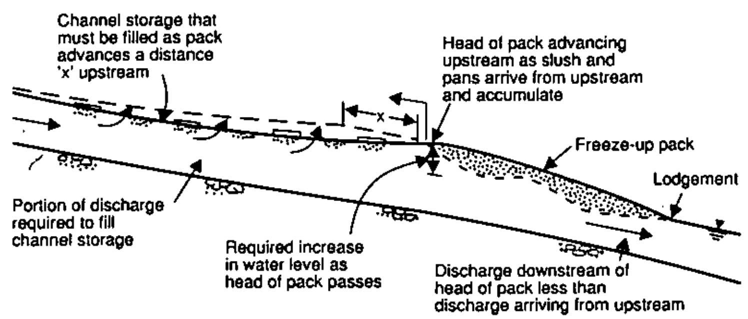


Figure 2.6 Loss of discharge to channel storage during progression of an ice cover (from Gerard, 1990)

Mid-winter flows are usually relatively stable, following a classic smooth recession curve over winter as ice cover thickens and baseflow sources are depleted. Gradual and relatively slight changes in resistance and flow can occur due to decreasing cover-bottom roughness (see Section 2.3.2) or, where covers are not entirely free-floating, a progressive blockage of the flow depth by ice growth. Such blockage can cause overflow onto the flood plain even at anomalously low flows (e.g., Forbes, 1979). The retention of water through storage and ice growth can produce a substantial flow when released in a relatively short period during the spring break-up.

The dynamics of streamflow during break-up and the role of ice in that process have received the most attention of all aspects of winter hydrology (e.g., Michel, 1971; Beltaos, 1983; Calkins, 1983, 1986; Gerard, 1988; Gerard, 1990). For many ice-covered rivers, the largest disruption to stage-discharge relationships and the greatest flooding occur because of the backwater effect created by break-up and ice jam processes. Notably, such ice-induced flooding often exceeds that which occurs under open-water conditions and at significantly higher discharge. Hydrologic analysis of the frequency of winter stage and discharge data has also shown that distinct populations of peak flows exist under ice-covered and open-water conditions (Gerard and Calkins, 1984). In general, the severity of break-up and its effect on the hydrograph falls into a continuum of conditions between two break-up extremes, commonly referred to as a) premature, mechanical, or dynamic and b) overmature or thermal (see Gray and Prowse, 1993).

In the thermal case, break-up conditions are comparable to those on a lake. The ice cover is in such a weakened state at the time of spring freshet that it offers only minimal resistance to break-up and, hence, water levels become only slightly elevated. Even ice jams produced by such conditions are usually relatively thin and rarely produce flooding (e.g., Prowse *et al.*, 1986). By contrast, dynamic break-ups are characterized by highly elevated water levels produced by a strong ice sheet resisting a large spring flood wave. Resultant ice jams produced by such break-ups often cause extensive flooding. Break-up is frequently characterized by a series of surges and stalls related to the formation and release of ice jams (Figure 2.7). Notably, as an ice cover is broken during the advance, water is effectively released from storage and can contribute to the flow driving the break-up progression. Melt of the ice cover can also significantly contribute to the flow volume during ice-jam decay as discussed in Section 2.8.6.

The rapidly varying flow and ice conditions, and the exceedingly high velocities (see Section 2.5.7) make conventional velocity/discharge measurements impractical or even impossible during ice runs and jams. Although alternative methods exist, as described in subsequent Section 2.4.6, these are rarely employed in regular monitoring programs.

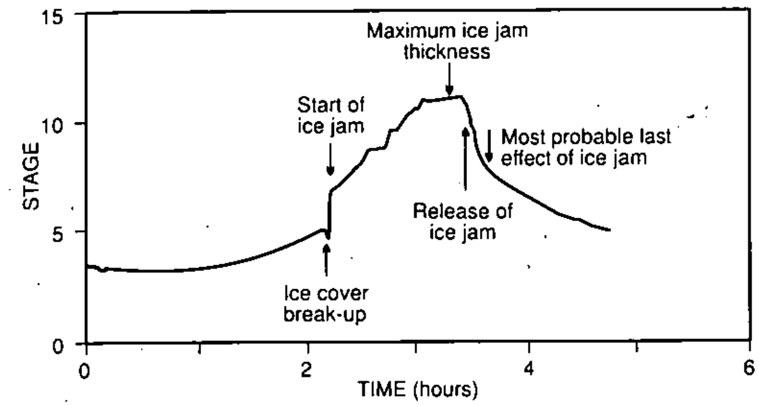


Figure 2.7 A typical break-up hydrograph (after Calkins, 1986)

2.4.3 Low Flow

During winter most precipitation falls in the form of snow and goes into seasonal storage thereby reducing landscape runoff. Channel flow is then largely sustained by groundwater and/or lake outflow which gradually become depleted over the winter season, typically resulting in the lowest flows late in the winter and prior to break-up. Superimposed on this basic relationship between winter and low-flow hydrology are several features of an ice regime that directly affect flows. In most instances, these ice factors further reduce flows and accentuate any low-flow problems that exist.

As outlined in Section 2.4.2, storage effects (Section 2.3.2) of an ice cover can shift the period of minimal low flow from late-winter to the fall freeze-up. This latter period is also when ice growth is most rapid, leading to a further abstraction of water from the flow and enhancing the low-flow condition. Such abstraction continues throughout the winter, decreasing in accordance with the rate of ice growth, but is often still detectable even a short time before spring melt and subsequent break-up (e.g., Sherstone, 1985; Blachut, 1988). Assuming temperature to be a good index of heat loss and hence ice growth, Gerard (1981) determined a relationship between degree days of frost and the amount of flow abstraction for a normal, instream ice cover.

The effect of ice-abstraction on flow is expected to be most pronounced where the ice-cover area and thickness are large relative to the channel width, depth, and flow rate.

Estimation of low-flows affected by ice covers also poses problems for regional analysis techniques (ASCE Task Committee on Low Flow Evaluation, 1980). Gerard (1981) investigated various catchment parameters to correlate with low flows, including unfrozen stream length, lake area in the catchment, and surficial geology. In a regional analysis of winter low flows, Melloh (1990) found that of various hydrologic and physiographic parameters, drainage area, latitude, and mean basin area correlated best with winter flows.

The coincidence of the annual low-flow period with the winter ice regime of a river has implications for many resource uses including domestic and industrial water supply, power production, effluent dilution, and overwintering aquatic habitat. If winter water balances or base flow volumes are being calculated for water licensing, effluent dilution, or availability of aquatic habitat, abstraction of flow due to ice storage and formation must be taken into account.

2.4.4 Precipitation Load Effects

Loading of an ice cover by precipitation can also modify the winter flow regime of rivers, particularly those fed by a large area of reservoirs and/or lakes. Accumulation of deep snow depresses the ice cover, displaces stored water and increases discharge. In a study of run-off in northern Quebec, for example, Jones (1969) found that up to 11% of the total winter run-off was created by such displacement in a catchment with a 23% lake area. Moreover, an average increase in lake discharge of almost 20% has been calculated by Kuusisto (1984) for 45 unregulated Finnish lakes, assuming only a 50 mm increase in the snow water-equivalent on these lakes. Such flow can be an important modifier of the winter flow regime, particularly in areas of low winter discharge.

2.4.5 Groundwater

The winter base-flow of a stream that is not lake-fed consists primarily of groundwater discharge (van Everdingen, 1988). The relationship between the ice cover of a river channel and the groundwater regime of the surrounding basin is, however, complex, poorly understood, and rarely studied in detail: e.g., Wankiewicz,

(1984a,b) being two of the few papers to grapple with the subject. In a small river or stream, the inflow of warm groundwater can also play a significant role in ice-cover development and deterioration. Its significance is more complex on large rivers, particularly in northern areas where the presence of permafrost creates confined groundwater aquifers or taliks (unfrozen zone) around a stream channel.

The characteristics of permafrost aquifers, which can be classified as suprapermafrost and intrapermafrost, are described in detail by van Everdingen (1987, 1990). A suprapermafrost aquifer is associated with a shallow closed talik surrounding a small stream which has a finite volume that varies seasonally, or a small lateral talik associated with a spring. This type of aquifer usually has a finite and small volume of flow available for discharge during the winter months, resulting in the development of a river ice cover in proportion with ambient climatic conditions. This type of stream is also conducive to freezing to the bed (Bates *et al.*, 1968). An intrapermafrost aquifer is found in a larger open talik through the permafrost, such as is found beneath a larger river. Hydrothermal or hydrochemical taliks are maintained by the transport of significant volumes of warm or mineralized water from the subpermafrost zone. Intrapermafrost aquifers are usually characterized by higher groundwater discharge-rates and larger volume discharges, resulting in more significant impacts on river ice regimes.

In general, the magnitude of the thermal contributions from groundwater, shallow or deep, the water flowing beneath stream channels, and the interaction with the ice cover have not been sufficiently established, yet this flow system provides the over-wintering environment for many biological species.

2.4.6 Streamflow Measurement

The effect of the presence of ice on streamflow and its measurement has been the subject of considerable analysis since the turn of the century (Barrows and Horton, 1907; Hoyt, 1913; Moore, 1957; Chin, 1966; Rosenberg and Pentland, 1966; Rantz *et al.*, 1982; Pelltier, 1988). Different techniques of streamflow measurement under ice conditions have been developed and refined over the years (Rantz *et al.*, 1982). A review of techniques and the methods employed in various countries is contained in Pelletier (1988, 1990).

In general, most estimates of flow during the ice-covered period rely on continuous measurements of stage. Seasonal variations in ice roughness (see Section 2.3.2), however, mean that the relationship between stage and discharge also varies through the winter. Standard practice is to conduct periodic flow measurements and, guided by the stage record, employ a suitable technique for interpolation between measurements. Different techniques apply to streams of different types, climatic regimes and even different seasons on the same stream (Figure 2.8). Small streams and temperate climates create most problems, particularly in the case of intermittent ice covers, where an ice cover can form, melt out and reform numerous times over the winter.

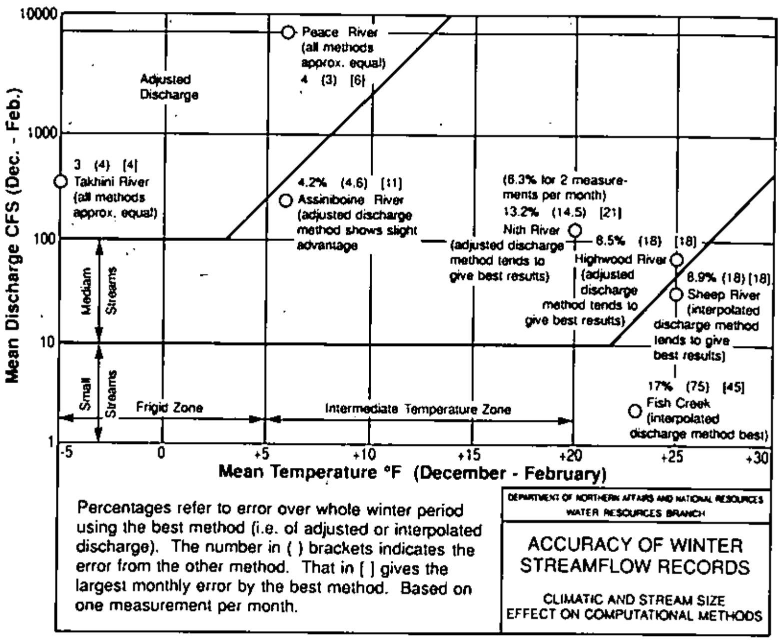


Figure 2.8 Dependence of errors in interpolated discharge during the winter season on climate, stream size and method (from Rosenberg and Pentland, 1966)

The rapidly varying flow conditions that characterize freeze-up and break-up are the most difficult to estimate using conventional techniques and, as a result, these are often the most unreliable periods of record in published data sets. This has special significance in the evaluation of

annual flows since break-up can also be the time of peak annual discharge, particularly on sub-arctic nival rivers (Church, 1974). Guidelines for extracting ice-break-up data from hydrometric station records are contained in Beltaos (1990) and for conducting detailed ice/flow measurement programs in Prowse (1990a). Other site-specific methods that rely on measuring velocity from shore-based and aerial surveys are reviewed by Prowse *et al.* (1986) and Prowse and Demuth (1991); methods that use sonar-ranging equipment for measuring ice and water levels from above the surface are reviewed by Gerard (1990).

2.5 TRANSPORT AND MIXING PROCESSES

2.5.1 Introduction

The presence of ice in a river significantly alters the transport and mixing characteristics of the flow. Such effects are most pronounced where the ice is in the form of a stationary cover that creates an additional flow boundary. Except for some infrequent and short-lived conditions, the ice cover, along with any snow load that may be present, is in a state of flotation so that it can generally move up or down with the water level. As in the open-water case, the flow is essentially driven by gravity.

The effects of an ice cover on transport and mixing processes are considered in this section. A comprehensive, quantitative treatment of riverine transport and mixing is beyond the scope of this report, but there are several texts and articles containing detailed information: e.g., Henderson, 1966; Yalin, 1972; Ashton, 1986; Fischer *et al.*, 1979; Beltaos, 1978; Elhadi *et al.*, 1984. Instead, emphasis is placed on the underlying physical mechanisms so as to illustrate the effects of the ice cover and identify knowledge gaps that require investigation.

2.5.2 Velocity and Shear Stress

For open-water flow in a very wide rectangular channel, the well-known logarithmic and linear distributions respectively describe the velocity and shear stress (Figure 2.9). The average velocity, V, and the bottom shear stress, τ_o, are related by a flow resistance equation: i.e.,

$$\tau_o = (f / 8)\rho_w V^2 \qquad (2.7)$$

where f = friction factor of the bed which, for "fully rough turbulent" flow, as is the usual case in natural streams, increases with relative roughness. The latter is defined as the ratio of absolute roughness, K_b, to flow depth, Y_o.

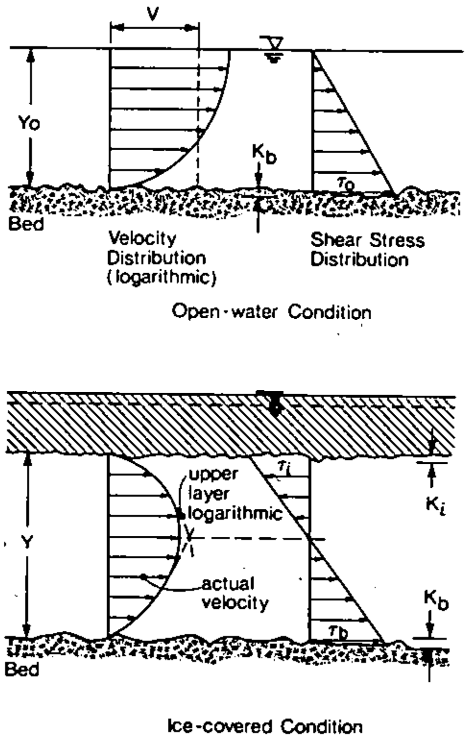


Figure 2.9 Velocity and shear stress distribution in uniform, unidirectional flows

For ice-covered flow, the shear stress remains linear, becoming zero at a depth dictated by the two roughnesses K_i (ice cover) and K_b (bed). The flow velocity is zero at the top and bottom of the profile and has a maximum at a depth that nearly coincides with that of zero shear stress. (Lack of symmetry, occurring when K_i is not equal to K_b, is the cause of non-coincidence between these two points: see, e.g., Hanjalic and Launder, 1972.) A common, though only partially correct, approximation, as used in Section 2.3.2, is to view the composite, ice-covered flow as a superposition of two open-water layers, the upper one being inverted (Figure 2.9; a modification of the more simplified Figure 2.1). Using this concept, simple equations can be derived to calculate the shear stresses τ_i and τ_b , and the average velocity of the composite flow. It should be kept in mind that we are discussing flow in a straight, very wide channel of constant depth. In natural streams, where various irregularities and bends are present, vertical velocity distributions are occasionally very different from the idealized forms of Figure 2.9. Consequently, flow resistance equations in both open-water and ice-covered river flows are only meaningful in a reach-averaged sense.

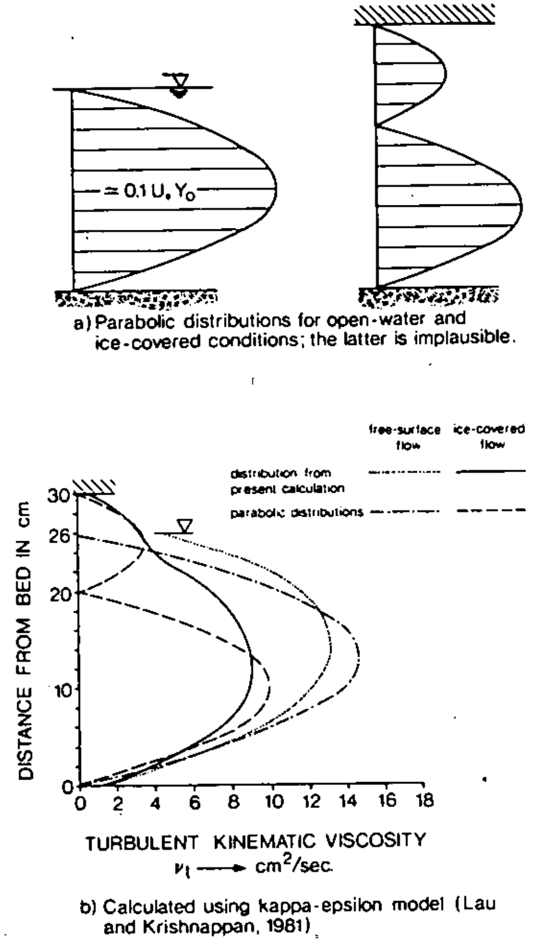
The additional resistance to flow caused by the ice cover reduces the velocity so that to pass the same discharge, an ice-covered, unregulated channel has to be deeper than when it is open. As discussed in Section 2.3.2, the increase in flow depth depends, among other things, on the roughness of the ice cover. If this roughness is equal to that of the bed, then the increase in depth will be ~30%, but it can be much more severe (100% or more) under an ice jam, i.e., a porous and characteristically rough accumulation of ice floes. The water level must rise to accommodate also some nine-tenths of the jam's thickness, possibly several metres, which explains the frequent flooding caused by ice jams. As noted earlier, this is the main consequence for which ice jams are known. They do, however, cause additional problems, as discussed in subsequent sections.

2.5.3 Vertical Diffusivity

In rivers, the main diffusive mechanism responsible for vertical spreading of dissolved or suspended substances is the turbulence of the flow. It is commonly assumed that momentum and mass transfer processes are analogous, so that the vertical diffusivity, E_y, is calculated from:

$$E_y = (\tau / \rho_w) / (du / dy) \qquad (2.8)$$

in which τ = shear stress at depth y; u = velocity at depth y; and du/dy = vertical gradient of u at depth y. (Equation [2.8] describes a parabola (Figure 2.10a), so that the depth-averaged diffusivity E_y amounts to $0.07u^*Y_o$ ($u^* = \mathrm{sqrt}(\tau_o/\rho_w)$, where u* is the shear velocity).



a) Parabolic distributions for open-water and ice-covered conditions; the latter is implausible.

b) Calculated using kappa-epsilon model (Lau and Krishnappan, 1981)

Figure 2.10 Vertical diffusivity distributions in uniform, unidirectional flows

This concept breaks down when applied to the two-layer, ice-covered flow representation because it gives an unrealistic distribution of E_y (Figure 2.10a); for example, it requires $E_y = 0$ at the point of zero τ,

implying no transfer of mass across this plane. Figure 2.10b, reproduced from Lau and Krishnappan (1981), shows distributions of E_y calculated using the "kappa-epsilon" turbulent flow model. It is clear that E_y does not vanish near mid-depth under an ice cover. On the whole, the vertical diffusive capacity of a stream is reduced in the presence of an ice cover.

2.5.4 Transverse Mixing

In purely two-dimensional, open-channel flow, the transverse spreading of a pollutant is mostly accomplished by turbulence, the corresponding diffusivity being proportional to u^*Y_o. In natural streams, however, additional spreading mechanisms are at work, caused by secondary currents. These arise from vertical and transverse components of velocity so that individual fluid particles move along helical paths.

There are two kinds of secondary currents. The first kind is produced by an imbalance in the normal Reynolds stresses, in turn caused by the fact that turbulence is not identical in all directions. ("Normal Reynolds stresses" are time-averages or ensemble-averages of squared velocity fluctuations, multiplied by the fluid density.) Such currents are normally very weak but may become significant if the aspect ratio of the stream is less than about 10.

River bends give rise to a much stronger type of secondary current, driven by radial accelerations. Figure 2.11 illustrates this kind of circulation for both open-water and ice-covered flows. In the latter case, two "cells" are present, consistent with the two-layer, composite flow concept. Measurements (Zufelt, 1988; Urroz, 1988) indicate that the magnitude of the radial (transverse) velocity component, relative to the tangential (longitudinal) component, is about the same for both open-water and ice-covered conditions (~0.1). Because the radial velocity distribution is not uniform, a dissolved or suspended contaminant would be radially advected at different rates, depending on its vertical position. This "differential advection" causes "dispersion", a term that has been adopted to describe spreading of aquatic substances due to non-diffusive mechanisms.

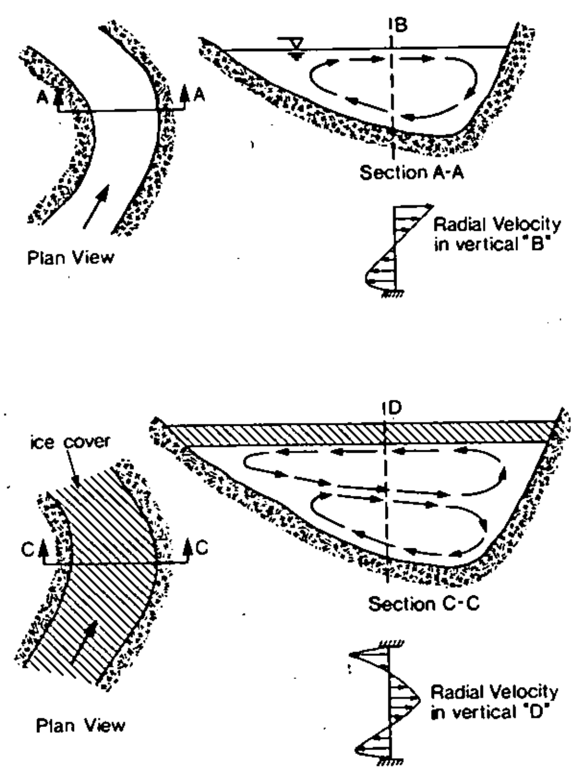


Figure 2.11 Schematic illustration of secondary bend flow in open and ice-covered channels

Transverse dispersion is commonly quantified by using an augmented or apparent diffusivity, called the transverse mixing coefficient. This assumption, by and large empirically based, greatly simplified mixing calculations by focusing on the depth-averaged concentration. The transverse mixing coefficient, E_z, is obtained by comparing measurements with solutions of the depth-averaged continuity equation for the spreading substance. Field data indicate that E_z scales on shear velocity, u*, and flow depth, Y, for both open-water and ice-covered streams. (Note u* is taken as (sqrt(gRS)) in which g = acceleration of gravity and R approximates Y for open-water conditions or $Y_i/2$ for ice-covered conditions; see equation [2.3]) It is not known why this is so, as very little work has been done to elucidate the effects of secondary currents on transverse mixing. An empirical relationship between K_z ($=E_z/u^*Y$) and river sinuosity has been obtained

by Lau (1985) and can be used to estimate E_z (Figure 2.12). Note that for straight channels, K_z has the well-known value of ~0.2 but increases to ~1.0 in a stream of sinuosity = 1.3. The large scatter of the data points in Figure 2.12 suggests that additional factors may have an influence on E_z.

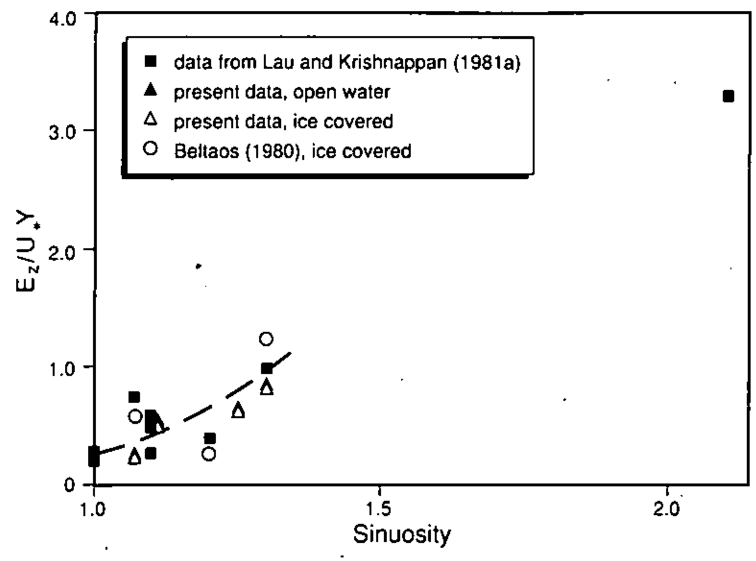


Figure 2.12 Dimensionless transverse mixing coefficients versus river sinuosity (after Lau, 1985)

Recently, Pavlovic and Rodi (1985) and Demuren and Rodi (1986) have further demonstrated the mixing ability of secondary currents through numerical modelling in two and three dimensions.

2.5.5 Longitudinal Dispersion

As discussed earlier, the term dispersion is commonly applied to the spreading process caused by differential advection. Longitudinal dispersion by far exceeds longitudinal turbulent diffusion in rivers and is essentially caused by transverse variations of the longitudinal velocity (Fischer *et al.*, 1979). The lack of prismaticity, i.e., variations in cross-sectional properties along the river, further enhances the dispersive capacity of natural streams (Beltaos, 1980; Li, 1989). Very little experimental information exists with regard to the effects of an ice cover on longitudinal dispersion. Based on the current understanding of the relevant factors, and invoking the two-layer flow concept, one would expect that the longitudinal dispersion capability is reduced by the presence of an ice cover. However, test data are needed to check this extrapolation. Since the time of writing, several field test results have become available. They are reviewed, together with implications to mathematical modelling of dispersion, in Beltaos (1993b).

2.5.6 Sediment Transport

Many aspects of the transport of sediment by rivers have been elucidated in the past few decades (e.g., see Yalin, 1972) and the subject is still under active research. Nearly all of this work relates to open-water conditions. Once the bed shear stress exceeds a threshold value, the bed sediment begins to move as "bed load" and "suspended load". The bed load is the amount of bed material being transported per unit time in close proximity to the bed. It is strongly dependent on the bed shear stress as well as on sediment and fluid properties. The suspended load is material that is transported mostly in suspension, with only infrequent contact with the bed. Vertical diffusivity and flow velocity are important factors in determining the suspended load, in addition to bed shear and sediment/fluid properties.

For the same discharge, the presence of an ice cover generally reduces the sediment "driving" parameters (shear stress, velocity, diffusivity), so the sediment transporting capacity should be significantly reduced. This expectation has been confirmed by Lau and Krishnappan (1985). Their experiments and analysis supported use of the two-layer concept of ice-covered flow, provided the vertical diffusivity is evaluated realistically; e.g., see Figure 2.10b. More experimental data are needed, however, both in the laboratory and in the field to define fully the effect of an ice cover. Existing laboratory studies include, for example, work conducted by Sayre and Song (1979) and Wuebben (1986; 1988). Field measurements are almost non-existent except for some work in Alaska on the Tanana River by Lawson *et al.* (1986).

The preceding considerations apply to relatively coarse sediments that are not subject to formation of cohesive bonds and consequent flocculation. Relatively little is known about the behaviour of fine cohesive sediments (size <62 μ) in open-water flow (Partheniades,

1986). When it comes to flow under ice, the writer is not aware of any investigations or data. A major environmental effect of fine sediments is their strong adsorptive tendency so that the fate of many aquatic contaminants is governed by the transport, erosion and deposition of fine particles. Clearly, a sustained research effort is needed to understand these processes and how they are influenced by flocculation.

Sediment can also become entrained into an ice cover. This results from frazil entrapment of individual sand and gravel particles, from anchor ice releasing off the substrate and carrying material into the surface ice cover, or from the ice cover freezing into the bed with subsequent dislodgement of ice and bed material by river-stage rises.

2.5.7 Scour

Riverbed scour occurs wherever the incoming amount of sediment is exceeded by the outgoing, and is generally divided into two categories, "general" and "local" scour.

General scour occurs over a large portion of the river width and length and is caused by gradual variations in flow characteristics. For example, general scour occurs at a constriction because of the longitudinally increasing flow velocities and shear stresses. The opposite is true downstream of the constriction so that we would expect deposition to take place. Local scour results from localized and highly variable flow types such as jets and vortices created by obstacles or artificial boundary configurations. For example, local scour occurs near bridge piers due to vortices forming as a result of the three-dimensional configuration of the flow around the pier.

Ice jams are the main scour-causing ice-cover type. Non-uniformities in the longitudinal configuration of a jam cause longitudinal gradients of the sediment transport capacity, thus leading to general scour and deposition regions as shown in Figure 2.13 (e.g., see Mercer and Cooper, 1977; Wuebben, 1988). Of course, the occurrence of scour presupposes that flow velocities and thence bed shear stresses are strong enough to exceed the "threshold" condition that may or may not be the case under an ice jam, depending on bed material size and flow discharge.

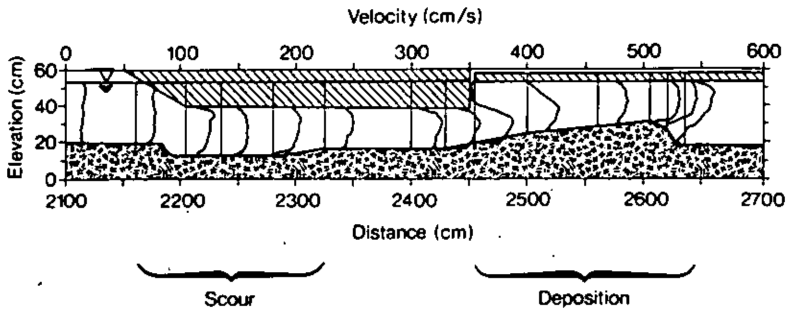


Figure 2.13 Scour-deposition patterns under a simulated ice jam (from laboratory tests by J. Wuebben, 1988)

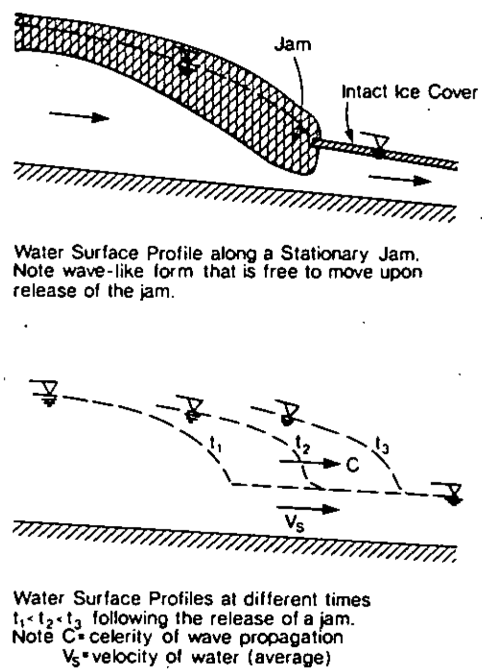


Figure 2.14 Schematic illustration of surges due to ice jam releases

The surges that accompany the release of ice jams (Figure 2.14) are known to propagate at very high celerities, C, and create unusually high water velocities, V_s (Henderson and Gerard, 1981; Beltaos and Krishnappan, 1982). Values of 5 m/s for V_s are not uncommon. For

a friction factor of 0.05, this value of V_s translates to a bed shear stress of 150 Pa, which, in turn, implies that the flow can move bed material particles as large as 20 cm in diameter.

Such flow conditions, though transient and perhaps not causing a large amount of general scour, will have two effects: (a) they will generally set the river bed in motion that may be detrimental to aquatic life attached to the bed, and (b) they will cause considerable local scour that may be detrimental to the safety of structures in the river or at the river banks. For example, Ashton (1987) mentioned an instance where an entire island disappeared from a river during the ice break-up, an event thought to have been caused by surging water and ice. Doyle (1988) reported on cases where rip-rap was damaged by the same process.

Virtually no quantitative information exists on the scouring caused by river ice. Field data would require considerable expense to obtain but no major difficulties are foreseen, especially if impulse-radar systems are used.

2.5.8 Summary

The presence of an ice cover in a river alters the flow field by introducing a new boundary. The resulting changes are fairly well understood and can be quantified with judicious use of the two-layer, composite flow concept.

Mixing processes are largely governed by turbulence and advection, and a fair understanding exists of the effects of ice covers. Gaps in knowledge about the ice effect are related to corresponding gaps in knowledge about open-water flow. For example, the transverse mixing coefficients of ice-covered rivers are not fully predictable, largely because the dispersive influence of helical motions at bends is not fully known.

Present understanding of coarse sediment transport in open channels suggests that the ice cover should cause a reduction of the river's transporting capacity, but this needs more detailed verification than is presently available either in the laboratory or in the field. When it comes to transport of fine, cohesive sediments, known to carry various contaminants in a sorbed state, relatively little is known under open-water conditions and nothing about the effect of the ice.

Ice jams appear to have the most serious effects on flow and transport processes, not only with regard to frequent flooding but also with respect to general and local scour and potential damage to river structures. Very little information exists, however, for predicting the scour that can be caused by ice jams. Both laboratory and field investigations are needed.

2.6 EFFECTS OF ICE ON SHORELINES

2.6.1 Introduction

The ice-related scour and erosion processes described in Section 2.5 can produce unique shoreline and substrate characteristics as described here and Section 2.7. The shoreline is defined herein as the land surface from the river waterline to the crest of the bank that usually borders a river (Figure 2.15). The bank is often characterized by unstable and slumping sediments with associated falling vegetation. The crest is where the sloping bank surface meets what is usually a more gently sloping land surface that often has undisturbed vegetation and *in situ* soil. This shoreline zone is predominantly a bank but may also include a strip of the river bed that acts as a beach along the toe of a river bank when the river water level is low.

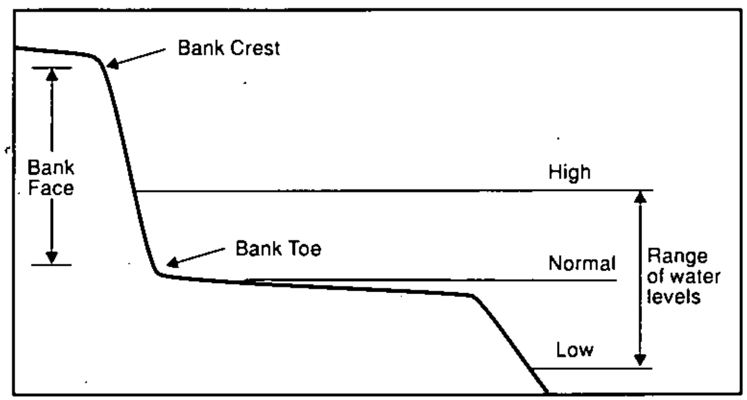


Figure 2.15 Idealized cross section of a shore zone (Gatto, 1988)

2.6.2 Ice Protection

Before addressing the specifics of how ice can erode shorelines, it is important to discuss the ice processes and conditions that protect shorelines from ice erosion. Shorefast and grounded ice can protect a river shoreline by limiting wave action and near-shore currents, and preventing the scraping, shoving and scouring caused by moving ice (Ouellet and Baird, 1978; Outhet, 1974; Hollingshead and Rundquist, 1977; Kellerhals and Church, 1980). This same protection can be provided by an ice coating or build-up on a bank face from onshore groundwater seepage (Carey, 1973) or from frozen overland sheet flow.

Ice pans (cakes) forced onto a bank by an ice jam can also provide protection to that bank as other ice pans are deflected or ride up on top of the first pan (Hollingshead and Rundquist, 1977). MacKay *et al.* (1974), for example, observed that ice shoved onto banks along the MacKenzie River often protected those banks from subsequent ice shove or pressure. When such stranded ice melts in place, however, a variety of irregular depressions, slump heaps and mud mounds that modify the river banks are often produced (Mackay and MacKay, 1977); the sediment released from the melting ice is washed downslope and helps to infill depressions and embed boulders.

2.6.3 River Ice Effects on River Shoreline Morphology

River shoreline (bank) conditions range from those that are stable with an intact vegetative cover and no apparent evidence of bank erosion or sediment instability, to those with features that characterize eroding banks or unstable slopes, including sediment flows, slides and slumps: "cleaned" banks with neither loosened sediments along the bank face nor sediment accumulations at the bank toe, sediment accumulations and pile ups, damaged vegetation, and erosion scars. River ice can contribute to or be solely responsible for the formation of any or all of these features. Bird (1967) suggests that the long-term geomorphic influence of river ice is difficult to assess and may be slight along many rivers.

The degree to which river ice processes contribute to eroding or unstable banks depends on the interplay of such factors as: river water levels and their fluctuations; ice strength, pressures and mobility; the degree of ice attachment to shoreline sediments and vegetation; the strength of shoreline sediments; and shoreline configuration. The first four factors are directly related to climate and shoreline sediment and vegetation conditions, while shoreline configuration is not.

Along a straight reach of shoreline, ice erosion would likely be more evenly distributed, with no location being eroded more than any other. Along an irregular reach, however, ice would be more likely to erode the banks on promontories than along backwater areas. Ice in the protected backwater areas would tend to be more stable than that along land jutting into the main part of a river (Gatto, 1984).

Along a river shoreline with a gently sloping near-shore bed, it is more likely that ice could gouge (MacKay *et al.*, 1974) or become attached to the near-shore sediment and never reach the banks until break-up, when river water levels are rising. Along a shoreline with a steep near-shore bed slope, ice can advance to the shoreline and erode them directly (Gatto, 1984).

2.6.4 Shorelines in Winter

2.6.4.1 Climate

Regional climates partially control river water level and its fluctuations (river discharge), and ice strengths (hydro-meterological conditions and modes of ice formation), which in turn affect ice pressures and mobility and the degree of ice attachment to a shoreline. Conditions characteristic of different regions vary enough that overlap between regions is considerable, and within-region, year-to-year variations can be substantial. Thus, only generalizations differentiate the regions (Figures 2.16 to 2.18).

In arctic regions, the warmest month has a mean air temperature of below 10°C and an average annual temperature less than or equal to 0°C. Arctic rivers are usually winter dry, permafrost often occurs at shallow depth beneath and along the rivers, and the spring snowmelt flood is apt to have the highest water level (Church, 1976). Information on ice strength and characteristics is not presented here but is adequately covered elsewhere: e.g., see Ashton, 1986; Prowse *et al.*, 1990. In general, variations in strength can be expected to occur in all ice-covered climates depending on the micro-meteorological conditions, particularly during the pre-break-up period.

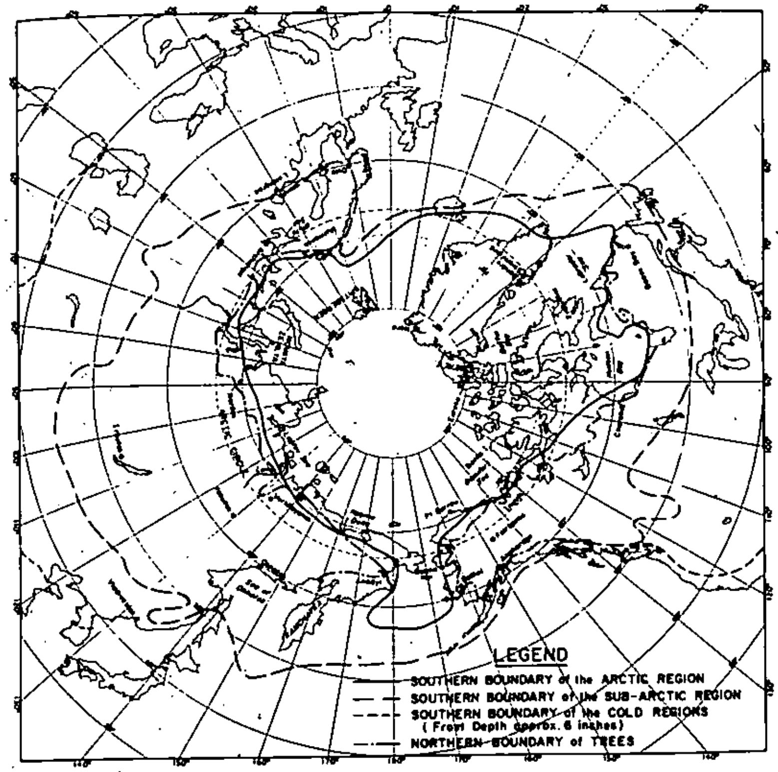


Figure 2.16 North Cold Regions: polar limits and zones (Stearns 1963)

In the subarctic region, the mean of the warmest month is above 10°C and of the coldest month below 0°C, and less than four months have a mean temperature above 10°C. Subarctic rivers experience generally low summer flows punctuated by periodic rainstorm floods that may exceed the spring freshet, especially on rivers with mountainous headwaters (Church, 1976). In a cold region, the air temperature is low enough, long enough to allow the seasonal frost penetration into the soil to reach at least 15 cm depth (Stearns, 1963).

2.6.4.2 Sediments and Bedrock

The distribution of frozen ground in the winter or year-round in some permafrost areas is related to the regions defined above. Considerable literature exists about soil processes and conditions in

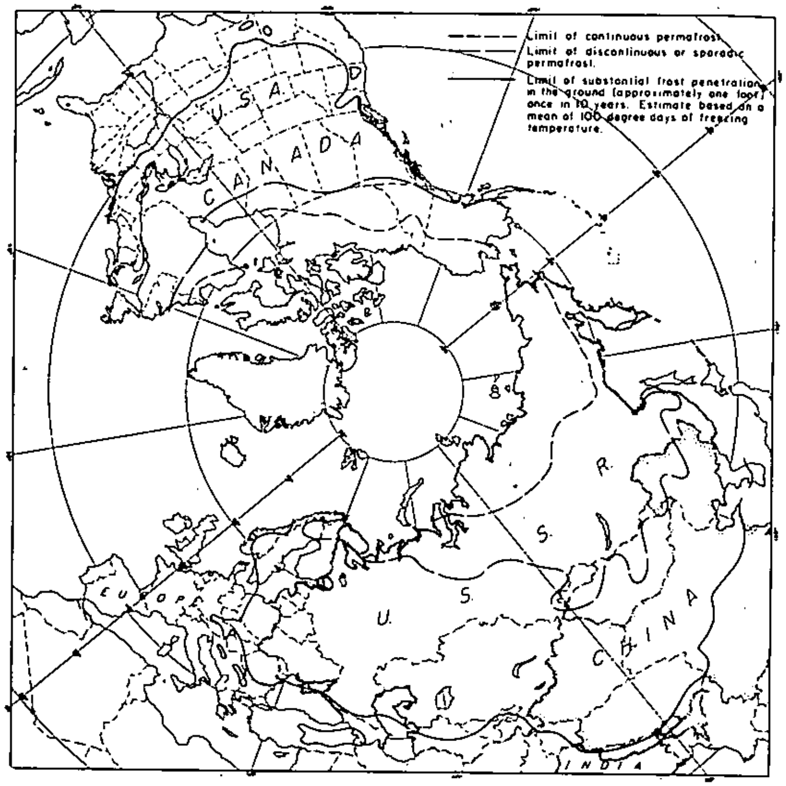


Figure 2.17 Cold region boundaries as determined by frozen ground (Bates and Bilello, 1966)

permafrost and seasonally frozen regions. Aspects of these are addressed here only to define the special conditions and processes that are present when river banks are frozen and thawed, when ground is perennially or seasonally frozen, and when river ice is present: and to explain the contribution of frozen ground, and freezing and thawing, to the complex of processes and conditions that affect river ice's interaction with a river shoreline.

Along bedrock shorelines, moving ice may scratch and polish soft rocks such as shales and limestones and dislodge large fragments of rock previously broken along joints (Dionne, 1974). Bird (1967) reports that irregular striae are produced on rock surfaces when river ice

containing boulders is pushed across them, but the total abrasion is slight; river ice polishes bedrock projections, but this has only local importance.

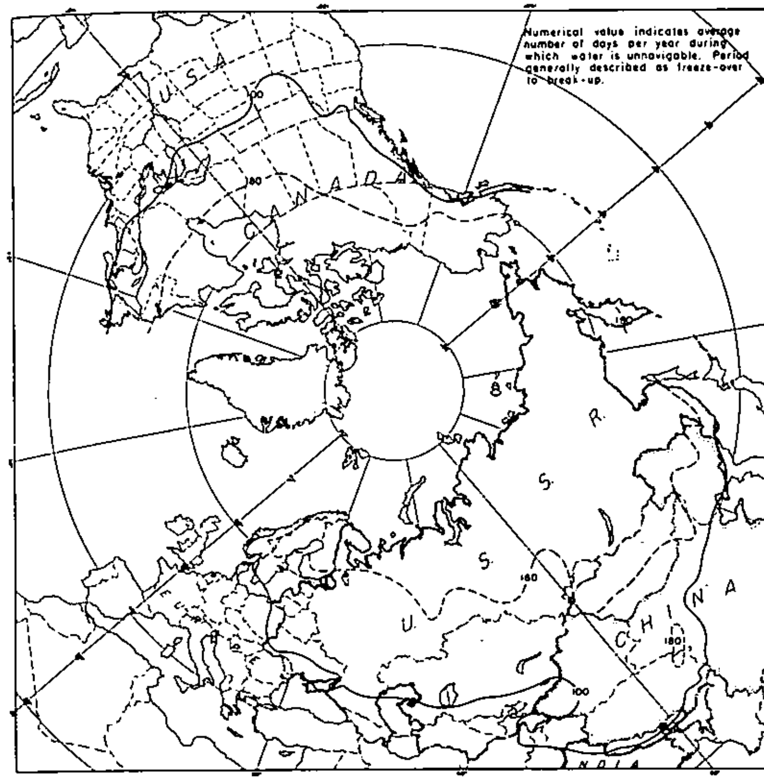


Figure 2.18 Cold region boundaries as determined by days with ice cover (Bates and Bilello, 1966)

These processes and conditions can increase the susceptibility of shoreline sediments to river ice actions: freeze-thaw (Farouki, 1981); groundwater piping (Hagerty, 1991a,b; Ullrich *et al.*, 1986); positive pore-water pressure; overland flows, including sheet and rill or gully (Brunsden and Kesel, 1973); water waves (often reduced in winter) and currents (Lawson, 1985); and water level fluctuations. Ice jams can cause spring floods which can cause the most severe erosion along northern rivers (Church, 1976).

The freezing temperature of soils varies through several degrees depending on their mineral, organic and water contents; and the frost penetration into the ground is directly related to air-temperature regimes complicated by many factors that affect surface heat losses and thermal conductivity, including the structural and textural composition of the soil; the insulation effect of vegetation and snow cover; the frequency, amplitude and duration of freeze-thaw cycles; the capacity of the soil surface for absorbing and retaining solar radiation; and the infiltration of rain and melt water (Gerdel, 1969). Therefore, to use a freezing index (e.g., the number of degree-days between the highest and lowest points on a curve of cumulative degree-days versus time for the average freezing season where the number of degree-days in any one day is the difference between the mean temperature for that day and the reference base of 0°C) to forecast frost depth in a region of seasonal freezing requires a thorough understanding of the local environment. In general, for a soil of uniform texture and moderate moisture content, frost will penetrate about 0.3 m for a freezing index of 56 (100 based on Fahrenheit degree days), and 1.8 to 2.4 m where the freezing index is approximately 2800 (5000 based on Fahrenheit freezing index) (Gerdel, 1969).

If shoreline or bank sediment is frozen, ice effects can be minimal, depending on the extent to which the sediment is frozen and the amount of ground ice present. Ice erosion along river channels of the outer Mackenzie Delta, for example, is not significant because banks are still frozen and thus more resistant when high river stages in the spring bring ice floes in contact with the shoreline banks during ice break-up. Bank erosion does occur at this location but primarily during the summer when water levels are lower (Hollingshead and Rundquist, 1977). It may be that river shorelines that are frozen during spring floods, when ice break-up and ice runs often occur, are generally changed only in minor ways compared to the extent to which they are affected by other processes once the ice has moved out. Data to confirm or disprove such a suggestion are sparse. Further discussion is provided in Section 2.7.

2.6.4.3 Vegetation

The northern limit of trees (start of the tundra) roughly coincides with the temperature boundary for the arctic region (Stearns, 1963). River shoreline vegetation in this region includes tundra grasses and shrubs (Figure 2.19). The southern boundary of the subarctic woodland is more difficult to establish and is not readily physically apparent

(Stearns, 1963). River shoreline vegetation here includes alders, willows, mixed forests, grasses and shrubs. In regions subject to seasonal frost shoreline vegetation is highly varied.

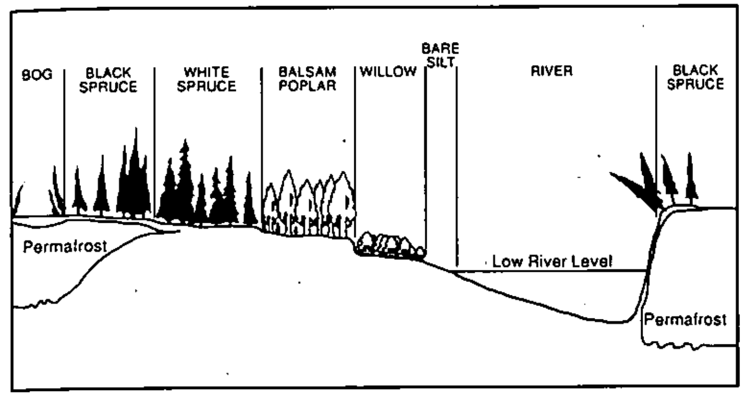


Figure 2.19 Diagrammatic cross-section of typical distribution of vegetation and permafrost across a meander of a river in interior Alaska (Viereck, 1970)

2.6.5 Shoreline Features Formed by River Ice Actions

Moving river ice can directly abrade (Bird, 1967), scour, or push *in situ* bank face sediments or vegetation, and can remove sediment or vegetation frozen into it by rafting when the river water level is high enough for the ice to act directly on the bank (Figure 2.20). The effectiveness of river ice as an erosive agent depends largely on whether break-up occurs on a rising or falling water stage. On a rising stage, shallower portions of the channel can accommodate the floating ice so that it may move downstream without much grounding and little bank erosion. On a falling stage, ice is likely to become stranded increasing the chances of bank erosion (Walker, 1969). Doyle (1988), for example, has described widespread bank erosion caused during an ice run on the Nicola and Coldwater Rivers and Spius Creek in southern British Columbia. Running ice also abraded riprapped banks, sometimes removing all the protective rocks, and seriously damaged floodplain trees, shearing some off.

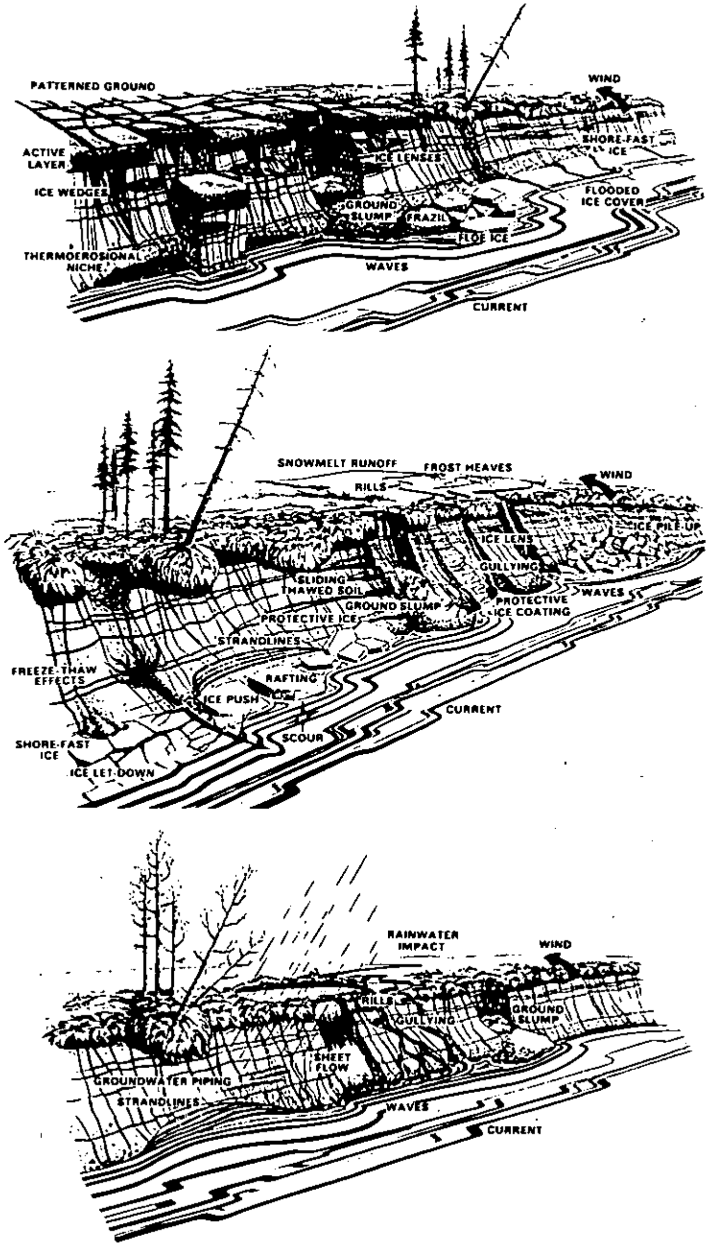


Figure 2.20 Shoreline features and processes - perennially frozen areas - seasonally frozen areas - unfrozen areas (Gatto, 1983)

However, Bird (1967) reports that erosion by floating ice is rarely a significant geomorphic process. If the water level is low enough that ice can only scour, push, and raft beach and near-shore bed sediment adjacent to the toe of a bank, the protection these sediments provides to the toe of the adjacent banks is reduced. The bank toe could then be more susceptible to erosion by waves and near-shore currents once ice has moved out.

River-ice erosion will most likely during spring break-up. It can cause local bank failures (Dionne, 1974) and may be more severe than scouring by water (Michel, 1970). However, Eardley (1938) reports that the amount of bank erosion by ice along the Yukon River appears to be inconsequential. Outhet (1974) report that bank scouring by river ice is less important in causing bank erosion than other processes.

Ice and river flow conditions are variable enough over many years along the same river that ice erosion along river shorelines can be important in one year and not the next. However, many of the damaging effects produced by river-ice erosion along shorelines during break-up, may be obliterated by subsequent water waves and currents as the water level reaches maximum spring high water (Koutaniemi, 1984). Mackay and MacKay (1977) report that the annual run of river ice in northern Canada produces a wide variety of ice-push features that are also observed in many cold fluvial environments of the world. Some of the features last a day, others last for centuries.

2.6.5.1 Sediment Slides

The disruptive effects that moving river ice may have on *in situ* sediment structure have not been addressed. Such ice shove could reduce sediment strength, which could lead to unstable sediments and slumping or to bank sediments with greater susceptibility to subsequent erosion by other processes.

2.6.5.2 Unprotected, "Cleaned" Banks

Shoreline sediment can freeze into or onto grounded ice. This ice can subsequently move and transport the sediment from its initial location (Dionne, 1974; Tsang and Szucs, 1972). Ice frozen to, or lodged against, river banks after spring high water can carry much fine bank sediment into the water when the ice finally falls down the bank (Bird, 1967). U.S. Army Corps of Engineers, Alaska District (1988) reported that bottom fast ice may carry only small amounts of sediments frozen to the bottom of the ice as it floats downstream. This rafting may be augmented by ice-scour and abrasion that lead to sediment removal. Removal can result in "cleaned" banks with no protective sediment accumulations at the bank toe or on the bank face. This lack of protection renders the bank vulnerable to direct attack from other processes.

2.6.5.3 Sediment and Debris Accumulations and Pileups

Banks of large northern rivers are characterized by numerous ice-push features such as the ice-shove ridges that top the banks of the MacKenzie River (MacKay *et al.*, 1974), or ice-push features such as the crescent-shaped ridges and boulder lines found along the banks of the Thomsen River, Banks Island. These latter forms are found where flow converges, presumably resulting in an accumulation of ice pans with subsequent push against the banks; observations show that ice activity is repetitive at these sites (Day and Anderson, 1976). Bird (1967) reports that permanent ridges may be pushed up by ice, but these rarely occur along North American rivers. He found them along the upstream sides of islands, as did Mackay and MacKay (1977) along Mackenzie River islands.

Other features resulting from ice-push include loose boulders and boulder pavements, buttresses, and debris piles (Mackay and MacKay, 1977; Dionne, 1974). At break-up, ice can be pushed up a river bank above the highest floodwater; this inland ice-push limit is often marked by a chaotic heap of driftwood, uprooted trees, mud and stone heaps (Mackay and MacKay, 1977). Bird (1967) reports boulder pavements along the Yukon River up to 500 m long and 30 m wide. On the Mackenzie River, they frequently extend to 7.5 m above the low water level and for many kilometres downstream. The pavements are composed of rounded and subrounded boulders washed out of till and pressed by ice into the silt and clay bottom and lower banks of rivers. They are also found on many other rivers, including the Thelon, Back and Dubawnt (Bird, 1967).

2.6.5.4 Erosion Scars

Ice-scour features such as gouges and ploughing tracks can be formed by moving ice that erodes shore sediment (Jahn, 1975; Outhet, 1974; Dionne, 1974; Shulits, 1972). Ice gouging and redistribution of bank material by iceshove are evident on many northern rivers, particularly along many reaches of the MacKenzie River

(MacKay *et al.*, 1974). Bank scour up to 15 m above summer base flow on the Mackenzie River was observed by Henock (1973). Measurements of ice-shove heights and driftwood heights revealed that the Liard River, Northwest Territories, Canada (a major northern tributary of the MacKenzie River) may rise as much as 10 m during spring flooding, and removal of bank materials by ice scouring may be a more significant process of erosion than water-undercutting/mass-movement mechanisms (Chyurlia, 1973). Further south, on the Red Deer River in Alberta, Smith (1979) observed erosion by blocks of floating and submerged ice during an ice run similar to that of a continuous line of bulldozers cutting into a hill slope. By contrast and still further south, the U.S. Army Corps of Engineers, Detroit District (1974) reported an absence of gouging of shorelines due to ice shoving along the St. Mary's River in Michigan and even suggested that shore ice formations served to protect against damage.

Kellerhals and Church (1980) report that the amount of morphological work due to ice-push and ice-scour appears to be insignificant in comparison to that of major floods with 5- to 10-year return periods. Their own field observations indicate that some scoured grooves along banks are usually no deeper than 10 cm and the gravel piles at the end of some grooves may be 20-50 cm high. Their review of previous work generally substantiated their findings that the role of river ice in channel modifications is neither significant in the long term nor is it widespread. Ice erosion can have local importance, however. King and Martini (1983) report that ice scours and ice-rafted materials greatly affect the river bank along the Attawapiskat River in the Hudson Bay Lowland. Ice floes gouge the river banks, introducing large amounts of fine sediment to the river in the spring, and coarse-grained, ice-rafted sediment deposits are common.

Water-current erosion scars and gouges can also be formed by water currents redirected by ice into the shoreline (Martinson, 1980). Icings and ice jams may even cause lateral migration of river flow by diverting water around the built-up ice into the shoreline (Hodel, 1986; Church, 1976).

2.6.5.5 Vegetation Damage

Vegetation that grows along river shorelines is not directly affected by river ice unless the ice scours it at the waterline or along the bank, or the vegetation is frozen to it and is subsequently dragged. When a river with an ice cover rises and falls, the ice can damage trees and shrubs as it recedes (Figure 2.21; Gatto, 1984). Scraping-off of bark is another common form of damage (Koutaniemi, 1984). Some of the most concentrated vegetation damage occurs at sites of ice jams and major ice pushes (Outhet, 1974; Reynolds, 1976). For example, ice thrusting along the Mackenzie River has been reported to have bulldozed over birch trees on a bank 7 m above the river (Mackay and MacKay, 1977). Chyurlia (1973) suggests it is very possible ice shoving or ice rafting during spring flooding along the Liard and Mackenzie Rivers are the principal agents of vegetation removal. Such removal can lead to the initiation of subsequent bank erosion by both collapse and scouring.

Figure 2.21 Trees broken by ice being let down as the reservoir water level dropped after a winter flood (December 1973, Franklin Falls Reservoir, New Hampshire) (Gatto, 1982)

2.6.6 Summary

Based on the current state of knowledge, the importance of river ice erosion along river shorelines or banks varies year to year and is generally considered to be not as important as water erosion and

associated sediment slides, although specific sites may suffer significant ice erosion in a particular year. There have not been sufficient studies to define more clearly the role of river ice in modifying the shore. Additional research in different geographic areas over several years is necessary to understand and define this role more completely. Research should also be conducted to evaluate the effects of river ice disruption of bank sediment structure and the possible increase in bank erodibility.

2.7 EFFECTS OF ICE ON SUBSTRATE

2.7.1 Introduction

As outlined in Section 2.6, processes associated with the formation, growth and break-up of river ice all can impact river shorelines to varying degrees. The same is true of the river substrate, although a physical separation of shoreline from substrate becomes debatable when ice-affected water levels submerge areas previously identified as shoreline under low-flow, open-water conditions.

This section first focuses on the effects of ice on major bedforms and sediment morphology of the substratum. Also discussed are specific substrate characteristics, including organic matter content and concentrations of nutrients and contaminants. A review of the effects of river ice on sediment temperature and permeability conclude the section.

2.7.2 Erosion and Deposition Effects

River ice, in its various evolutionary stages, is an effective agent of erosion and deposition on the substrate, particularly on alluvial rivers (e.g., Cook, 1967; MacKay *et al.*, 1974; Forbes, 1979; Smith, 1979; and Prowse, 1993). Although the majority of effects are most pronounced at break-up, ice-erosion of sediments can also occur during the initial stages of freeze-up. One erosional process, unique to freeze-up, is the scavenging of freeze-bonded sediments from the river bed by frazil and anchor ice. Although anchor ice can grow to quite large accumulations on heavy bed material, it rarely remains for any significant period on fine-grained sediments, simply because of the buoyancy effects of the anchor ice. The magnitude of such substrate erosion remains undocumented, but it may be quite large locally, particularly below turbulent reaches that favour frazil-ice growth.

Another erosional process related to frazil accumulations is the development of scour holes beneath aggrading hanging dams. Erosion of bed material is believed to occur because of enhanced flow velocities beneath or on either side of the flow-obstructing, frazil-ice accumulation (Majewski, 1990).

As suggested in Section 2.5.7, the most pronounced effects on the river substrate result from break-up processes, particularly those related to ice jams, runs and surges. The magnitude of such effects will also depend on the intensity of the event, ranging over a continuum from a premature, dynamic or mechanical break-up to that of an over-mature or thermal break-up. Erosion and subsequent deposition of materials should be highest during dynamic break-ups that are typically characterized by large spring flood waves and rapidly moving break-up fronts. In contrast, erosion should be least during thermal events where the ice cover weakens thermally to such a high degree that even the smallest spring flow can dislodge and remove it downstream.

Considerable physical evidence of break-up ice erosion has been collected for estuarine and river substrates including features such as ice drag scars, ice holes and kettle holes (Dionne, 1969; Day and Anderson, 1976; Mackay and MacKay, 1977; and, Gordon and Desplanque, 1983). Ice-drag scars, extending for considerable distances and eroding substantial substrate material, are typically the result of the downstream transport of grounded ice. For example, Day and Anderson (1976) identified old ice-scour grooves in the Thompson River, Banks Island, extending up to five metres in width and several hundred metres in length. Such scarring action is likely to be a common phenomenon associated with grounded ice jams, particularly where they form in shallow reaches, such as at river mouths, and over shoals, where ice runs frequently form large grounded ridges.

Where extensive grounding occurs, such as at the toe of some ice jams, scouring is likely to occur over large areas where high-velocity chutes develop in response to a concentration of flow and a deflection of the flow toward the bed (e.g., see laboratory experiments by Wuebben, 1988). Similarly, high-velocity scouring around and under grounded ice blocks is known to produce smaller erosional features, such as localized scour holes. Thawing of any blocks that become buried by break-up processes and later exposed by falling water levels, can also create kettle holes, such as the 2 to 3-m depressions found in sandy shoals along the Mackenzie River (Mackay and MacKay, 1977).

In addition to erosional features, river ice can produce distinct substrate forms through sorting and deposition, as discussed with respect to shoreline features (see Section 2.6.5). One of the most prominent features is that of boulder pavements formed by the regular gliding action of ice over bed material (e.g., Wentworth, 1932; Mackay and MacKay, 1977). Although most information about boulder pavements (see Scrimgeour *et al.*, 1993, for a review of similar boulder features) refers to the river bank, little is known about the extent or prevalence of such features within the main (open-water) flow channel where ice gliding action is also prevalent. Moreover, the importance of such unique features as a discrete habitat type for algae, invertebrates and fish species remains unknown, as discussed further in Chapter 4.

A less conspicuous depositional feature, but perhaps one more important to the river ecosystem, is the layer of sediment that forms as break-up activity diminishes. Unfortunately, records of peak suspended-sediment content are extremely rare for the active break-up period. In one study, however, Prowse (1993) documented break-up suspended sediment loads equivalent to that carried by 3 to 5 times greater discharge under open-water conditions. Such an observation suggests that a major component of the annual sediment cycle could be overlooked by conventional monitoring programs that do not normally include sampling during break-up.

Although sediment deposition within the main channel and in overbank flood zones is commonly observed after break-up, reports vary about its overall significance. For example, field data include observations ranging from that of a thin sediment-veneer to deep localized deposits over one metre in depth (e.g., Eardley, 1938). Such differences in the degree of post-break-up sedimentation are probably related to natural, spatial and temporal variations in the deposition process, which are driven by a diversity of factors, including annual variations in the severity of break-up and freeze-up processes (e.g., celerity, velocity, stage, ice strength), susceptibility of the channel to erosion (e.g., size and weight characteristics), and channel variations in velocity and morphology. For example, it would be intuitively expected that some of the greatest post-break-up deposits of sediment should occur in the backwater zone of a slowly releasing ice jam formed by a dynamic break-up on a fine-sediment alluvial river. Although a quasi-uniform deposition of sediments across the channel bed might be expected in such a case, jamming at meander bends can lead to enhanced deposition at the outside of the bend and erosion on the inside bank from flow diversion through the bend (e.g., Martinson, 1980).

Under open-water conditions, the reverse tends to be true. In extreme cases, such flow redirection can result in meander cut-off and can limit the sinuosity of a river (Williams and MacKay, 1973). Additionally, ice blocks are capable of transporting large amounts of sediments and are thought to play an important role in influencing the thickness of the sediment layers in estuarine ecosystems (e.g., Dionne, 1968a, b, 1969).

The combined effects of ice-induced erosion and deposition should also alter the size composition of substrate sediments. This has special significance for related substrate characteristics such as nutrients and contaminants, the binding of which is influenced by sediment particle size.

2.7.3 Substrate Characteristics

Scouring and deposition of river sediments, as well as lateral exchange between the river water and the flood plain, should alter some important chemical characteristics of river substrates, such as nutrient, organic matter and pollutant concentrations. For example, under the extreme scenario of dynamic break-up, ice scour can lead to a large input of organic matter from channel banks, floodplains, and upstream sediments. For rivers that are not deeply incised (i.e., unconstrained with broad valley floors), the formation of ice jams and the resultant inundation of the flood plain can also provide substantial input of dissolved and particulate organic matter (Mackenzie River Basins Committee, 1981). The quantity and quality of these materials are important factors, influencing the abundance of microbes, detritivorous invertebrates (i.e., those that use detritus as a food source) (Culp and Davies, 1985), and contributing to overall riverine productivity (Cummins *et al.*, 1983, Chapter 4).

Productivity in high-latitude rivers has been hypothesized to be limited by the availability of nutrients like phosphorus and nitrogen (e.g., Bodally *et al.*, 1989; Roy, 1989). Because dissolved and particulate organic matter inputs to rivers contain substantial amounts of phosphorus and nitrogen, river ice break-up likely has a major impact on substratum nutrient budgets. For example, peak nitrogen, phosphorus and extractable metal concentrations for the Athabasca River were found to occur in April and May during the period of break-up and maximum discharge (Mackenzie River Basin Committee, 1985). Information on the importance of ice-mediated inputs of nutrients to stream nutrient budgets is scarce.

Ice break-up also has important implications for spatial and temporal distribution patterns of pollutants in the substratum. For instance, downstream transport of polyaromatic hydrocarbons in the Mackenzie River is thought to be influenced by exchange rates between the water column and the sediments (Nagy *et al.*, 1987). Clearly, the spatial sorting of sediments from ice action, as described above, will alter exchange rates at the water-sediment interface. Moreover, break-up effects on the substrate should alter the large-scale distribution of pollutants. For instance, pollutants absorbed onto fine sediments at point sources during the low-flow winter period could be transported far downstream from their point of origin because of the enhanced transport of sediments during break-up (Prowse, 1993). Further discussion on the effects of river ice on water chemistry and the mechanisms involved is provided in Chapter 3.

In addition to affecting substrate bedforms and substrate chemistry, river ice can alter substrate temperature due to the production of frazil and anchor ice and development of surface ice (Walsh and Calkins, 1986; Tsang, 1982; Blachut, 1988). While anchor ice forms predominantly on coarse substrates and on objects protruding from the substrate, it can completely cover the stream bed (Tsang, 1982) and can occupy the entire space between the substrate and the surface ice (Michel, 1971). The development of anchor ice can be responsible for substrate freezing by causing stream flow to be diverted upwards breaking through the water surface. This flow allows frost to penetrate into the anchor ice, freezing it and the substrate (Walsh and Calkins, 1986).

The transition from open-water conditions to ice-cover conditions reduces the flux of short wave radiation through increased surface reflection and attenuation. This alters substrate temperature directly by reducing the amount of short-wave radiation received by substratum in shallow areas of rivers and indirectly by decreased warming of the water column. Finally, the presence of an ice cover on the substrate, and substrate freezing, can reduce interstitial stream flow and exchanges between streamflow and interstitial flow. By reducing exchange between the stream flow and interstitial flow, anchor ice could reduce dissolved oxygen concentrations in the substratum. Blachut (1988) listed these consequences of river ice formation as potentially important factors affecting the survival of chinook salmon eggs and alevins over the winter months in the Nechako River, British Columbia.

The presence of an ice cover on the substratum has important consequences for exchange of nutrients between the algal community and the water column, binding of pollutants with fine sediments, and uptake of pollutants by benthic dwelling organisms. Unfortunately, information on these impacts is scarce. Further discussion of the effects on river ice on the aquatic biota is presented in Chapter 4.

2.7.4 Summary

The above review indicates that the formation, growth and break-up of river ice are strong modifiers of substrate bedforms and morphology, and exert strong impacts on sediment temperature and permeability. In contrast to these known effects, considerably less is known of the impact of ice on sediment characteristics, including organic matter content and, nutrient and contaminant concentrations. Despite this lack of information, however, it seems likely that river ice should have strong impacts on the distribution and abundance of organic matter, nutrients, and contaminants. The magnitude of these effects, the underlying mechanisms involved, and the mechanisms influencing the effects of river ice on substrate bedforms and sediment morphology, all require further study.

2.8 THERMAL AND CLIMATOLOGIC EFFECTS

2.8.1 Introduction

The formation, growth and ablation of river-ice covers are all highly controlled by thermal and climatologic processes. For the most part, such processes are relatively well understood. However, the inverse effects of river ice on thermal and climatologic processes are less well understood. This chapter focuses on such effects. An introduction to the basic energy balances governing ice and open-water conditions in rivers is provided as background for discussing the relative importance of the various heat fluxes. Most of the discussion focuses on in-stream effects, such as water temperatures and underwater radiation-regimes, but some larger-scale climatic effects are also considered. The cited references do not form an exhaustive literature set but were selected to illustrate the scope of the thermal/climatologic environmental effects. In some cases, discussion of thermal aspects links directly with other related environmental aspects (e.g., biological habitats, Chapter 4).

Where this occurs only brief comments regarding the related significance is made and readers are referred to the other sections of this report for more detail.

2.8.2 Heat Budgets and Light Penetration

Prior to the formation of a river ice cover, water within the channel freely exchanges heat with its surroundings. A generalized heat budget for a section of open water can therefore be written as:

$$\phi_{\cdot o} = \phi_A + \phi_W \tag{2.9}$$

where $\phi_{\cdot o}$ is the total heat supplied to the water column, ϕ_A is the combined atmospheric heat flux, and ϕ_W are the heat fluxes from beneath and within the water column (positive values indicate a heat gain to the water column).

Under open water conditions, it is the atmospheric fluxes that are usually the greatest and consist of:

$$\phi_A = \phi{\downarrow}(1-\alpha_w) + \phi_L + \phi_H + \phi_E + \phi_P \tag{2.10}$$

where $\phi{\downarrow}$ is incident short-wave radiation, α_w is the albedo for water, ϕ_L is net long-wave radiation, ϕ_H is the sensible heat flux, ϕ_E is the latent heat flux, and ϕ_P is the precipitation heat flux.

The components of ϕ_W include:

$$\phi_W = \phi_R + \phi_B + \phi_F \tag{2.11}$$

where ϕ_R is the heat flux from surface tributary inflow including groundwater seepage, ϕ_B is the bed heat flux resulting from the release of sensible heat stored in the bed and the geothermal flux (the latter usually being about two orders of magnitude less than the former; Larsen *et al.*, 1986), and ϕ_F is the heat due to fluid friction.

Notably, a significant portion of heat stored within the bed is derived directly from short-wave radiation penetrating the water column during the open-water period, especially in the case of shallow rivers and streams.

Under open-water conditions, it is the atmospheric inputs that dominate the heat budget and control the temperature regime of river systems. Once an ice cover develops, however, the relative importance of the various fluxes changes dramatically. A generalized equation for the heat budget of ice-covered flow can be written as:

$$\phi_{*_i} = \phi\downarrow(1-\alpha_i)e^{-\nu t} + \phi_K + \phi_R + \phi_B + \phi_F \quad (2.12)$$

where ϕ_{*_i} is the heat available to warm the ice-covered flow, α_i is the albedo of the snow-ice surface, ν is a bulk extinction coefficient for the snow/ice, t is the snow/ice thickness, and ϕ_K is the water to ice heat flux.

The most notable changes in the available heat from open- to ice-covered conditions result from a decoupling of the convective exchanges (ϕ_H and ϕ_E) and modifications to the radiation regime. In the presence of an ice cover, atmospheric exchanges of sensible and latent heat occur at the air-ice boundary. Similarly, since incoming long-wave radiation is reduced to negligible levels within a few millimetres of the ice surface, the long-wave radiation balance between atmospheric and terrestrial radiation is determined at the same interface. Hence, the only effect of these three atmospheric fluxes on the water heat budget is through the modification of the vertical ice-temperature gradient and the related conductive heat flux. Strong cooling at the ice surface results in a steep gradient and hence high rates of upward heat conduction from the water surface, thereby creating potentially high rates of ice growth. The actual rate of ice growth, however, also depends on the hydrothermal heat flux (ϕ_K) to the water-ice boundary. Where ϕ_K exceeds the rate at which heat can be conducted upwards through the ice sheet, melting will occur. The hydrothermal flux largely depends on the water temperature, which is primarily controlled by the other terms in equation [2.12].

In the case of short-wave radiation, an ice cover reduces the flux through greater surface reflection and attenuation. An open water surface is relatively transparent to short-wave radiation, with a typical average (neglecting effects of incident angles) albedo (α_w) being less than 0.1; values for α_i are only as low as 0.1 for highly transparent columnar ice (Bolsenga, 1969). More commonly, they are much higher, in the range of approximately 0.3-0.5 for most granular ice forms, and as high as 0.9 for fresh new snow (e.g., see review by Prowse and Stephenson, 1986). In addition to being an effective reflector of short-wave radiation, ice can also strongly attenuate radiation. The relation between the percentage transmissivity, $D_{\%}$, and the extinction coefficient, ν, is:

$$D_{\%} = 100e^{-\nu t} \quad (2.13)$$

Typical values of ν include a near-minimum value of 2 for relatively transparent ice forms, approximately 4 for various white (snow-ice) forms, 10 for snow in a melting state, and a near-maximum of 40 for newly fallen, light-density snow (see Prowse and Stephenson, 1986 for review). Figure 2.22 displays the rate of attenuation for various water/ice types and thicknesses.

For relatively thin and/or transparent ice covers, significant amounts of short-wave radiation can still reach and warm the flow (and the bed in the case of shallow rivers). Such warming, however, varies according to daily and seasonal cycles in atmospheric short-wave radiation. When a snow cover is present on the ice cover, this term is close to zero.

Over the winter period, ϕ_B, ϕ_R and ϕ_F (for unregulated rivers) decrease as heat stored from the previous open-water season is released to the flow and as flow velocities decrease with lowering winter discharge, respectively. As spring approaches, ϕ_B is often exhausted but ϕ_F will increase with spring flow. More importantly, however, $\phi\downarrow$ also increases, not only due to the seasonal rise in incoming short-wave radiation but because of changes in factors controlling the snow-ice radiation regime. For example, as the snow surface metamorphoses and begins to melt-ripen, the albedo can drop from a near maximum of 0.9 to approximately 0.5. In the case of shallow snowcovers, reflection/refraction due to the underlying surface is likely to affect the surface albedo. This has been noted for shallow snow overlying ground (e.g., Gray and Landine, 1987), but the effect has not been documented for snow on freshwater ice. Given the myriad of ice cover types and their large range in optical properties, the effect is likely to be highly variable.

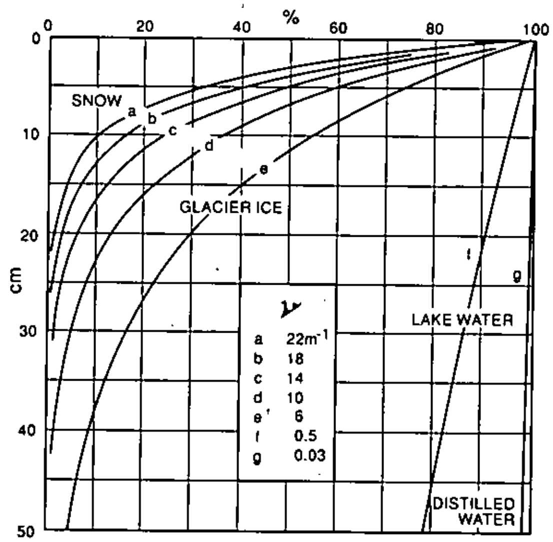


Figure 2.22 Penetration of light into snow, ice and water (after Geiger, 1961)

With the ablation of the snowcover and the exposure of the underlying, less-reflective ice layers, the albedo can further drop to as little as 0.1 in the case of transparent black ice. Overall, elimination of the overlying snow can produce an increase in available short-wave radiation by as much as 800%. Significantly, however, changes can also occur in the optical properties of the ice. Prowse and Marsh (1989), for example, note that the albedo of black ice increased to 0.52 due to an acicular melting process that eventually led to the granulation of crystal tops (Figure 2.23). To a large degree, such melt is produced by the absorption of short-wave radiation. It is not, however, limited to the surface of the ice sheet. Where internal portions of the ice have been warmed to 0°C, the attenuation of additional short-wave radiation will produce melt. Initially localized at grain intersections, small films of meltwater can rapidly grow into an interlocked network of channels and large pores. The effect of melt-porosity on the mechanical strength of an ice cover has been reasonably well documented but little is known

about its optical effect. The closest physical analogy is that of brine channels within sea ice, which are known to affect both the rate of short-wave radiation attenuation and surface reflection (e.g., Perovich and Grenfell, 1981). It is logical to assume that melt-pores will produce similar effects, particularly during the spring melt period.

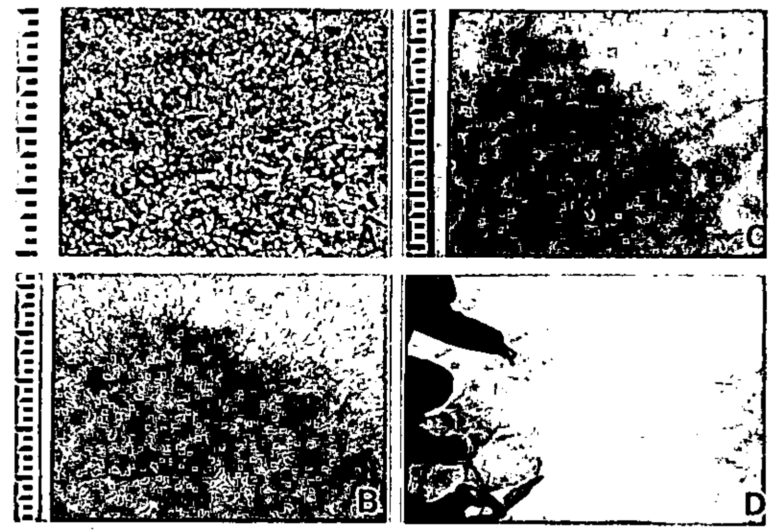


Figure 2.23 Progressive decay of black ice surface: (a) crystal pavement, (b) surface candling, (c) granulation of crystal tops, and (d) decaying black ice with appearance of a melting snow-ice surface (from Prowse and Marsh, 1989)

Common wisdom is to assume that radiation receipts at the water column are seasonally maximized once the snow cover is eliminated. Although this may be true immediately following the surface exposure of the ice sheet, subsequent melt-pore growth is likely to increase the reflective and attenuative losses, thereby reducing the amount of short-wave radiation reaching the underlying water.

2.8.3 Water Temperatures

Although temperatures beneath lake covers are known to climb to several degrees above freezing, under-ice water temperatures in a river rarely climb above a few hundredths of a degree, except in cases of significant inflow of warm water from tributaries or groundwater, or in streams where an air gap may exist between the ice and water surface. Turbulent flow conditions in a river promote rapid heat transfer to the ice (ϕ_K), thereby retaining the near-freezing temperatures. As described above, short-wave radiation can be the most significant heat-flux in warming the flow. However, because of its diurnal cycle, water temperatures can also vary between day and night (e.g., Marsh, 1990).

Some river biota, including fish, are known to have an under-ice preference in habitat related to water temperature. The degree to which this is controlled by the magnitude and timing of the various terms in equation [2.12] has not been investigated.

During periods of warm-weather that typically precede spring break-up, additions of heat to the water column ultimately lead to a warming of the ice sheet, thinning of the ice bottom (see Section 2.8.4), and the formation of ice ripples which further serve to enhance the water-ice hydrothermal heat transfer. For example, Ashton and Kennedy (1972) and Gilpin *et al.* (1980) found that heat transfer was 50 to 60% greater for rippled ice surfaces than for smooth ice. Hydrothermal warming of the cover also plays a role in decreasing the mechanical strength of an ice cover, a major factor leading to the eventual break-up of a river ice cover (e.g., see Prowse *et al.*, 1990).

As described elsewhere in this monograph, break-up is often a dramatic event capable of rapidly modifying physical and biological regimes, possibly providing the annual biological set-point for some river systems. The downstream progression of break-up and the accompanying open water front can also produce rapid changes in the overall thermal regime of rivers. Marsh and Prowse (1987), for example, measured water temperatures ranging from 3 to 5°C at the leading edge of open water associated with break-up on the Liard River near the confluence with the Mackenzie River. Even higher temperatures, in the 8 to 9°C range, have been reported at leading edges further downstream in the Mackenzie River (Parkinson, 1982; Terroux *et al.*, 1981). The downstream hydrothermal flux of heat is especially important on large northward flowing river systems because it acts as a long-range mechanism for conveying heat. Such heat is ultimately transferred by long-wave radiation and convective fluxes to the local climate. When such transfers occur during the spring period, with the surrounding landscape still snow-covered and relatively cold, their effects become especially important.

Water temperature at the leading edge of open water is largely controlled by meteorologic and break-up conditions (Prowse and Marsh, 1989). Where break-up proceeds sequentially downstream, such as in a "dynamic" or "mechanical" break-up, heating of upstream flow occurs largely unhampered by upstream ice effects. Water temperatures at the leading edge depend, therefore, on the prevailing open-water heat balance. The other extreme case is that of a "thermal" break-up, which often involves a non-sequential progression of break-up on northward flowing rivers. It is characterized by near-simultaneous break-up of the ice cover over extensive sections of a river, usually beginning at tributary mouths or in fast-flowing reaches. The end result is that ice and open-water sections are interspliced along a river. Hence, only limited warming occurs within an open water reach before it is cooled again beneath an ice-covered reach. Significant warming of the flow can only occur once the river becomes largely ice-free. This can be a protracted process, particularly where final ice clearance is hampered by low spring flows following the initial break-up.

2.8.4 Ice Cover Thinning

Cover thinning is a regular process that accompanies general warming. Figure 2.24 illustrates the rate of thinning reported for some northern rivers. Such decreases have a number of important implications including, for example, the bearing-strength and safety margins for over-ice transportation, under-ice discharge, and the overall severity of break-up ice runs and ice-jam-related flooding.

In general, cover thinning can occur at the surface from air-ice heat fluxes ($\phi\downarrow$, ϕ_L, ϕ_H, ϕ_E, ϕ_P) or at the bottom from the water-ice heat flux (ϕ_K). In a comparison of such fluxes, Prowse and Marsh (1989) estimate that under conditions of intensive warm air advection (large values of ϕ_H) and high solar radiation (clear sky radiation and low albedo), total atmospheric heat fluxes to an ice cover would be in the order of 20 MJ/m^2/d. This translates into a thinning rate of approximately 65 mm/d.

In the case of ϕ_K, it is more difficult to estimate extreme rates of ice thinning because changes in the cover and flow conditions can affect the rate of water-ice heat transfer. Essentially, the turbulent transfer of heat at the ice-water boundary is affected by the velocity distribution, which is determined by the ice and bed roughness, and the depth of flow.

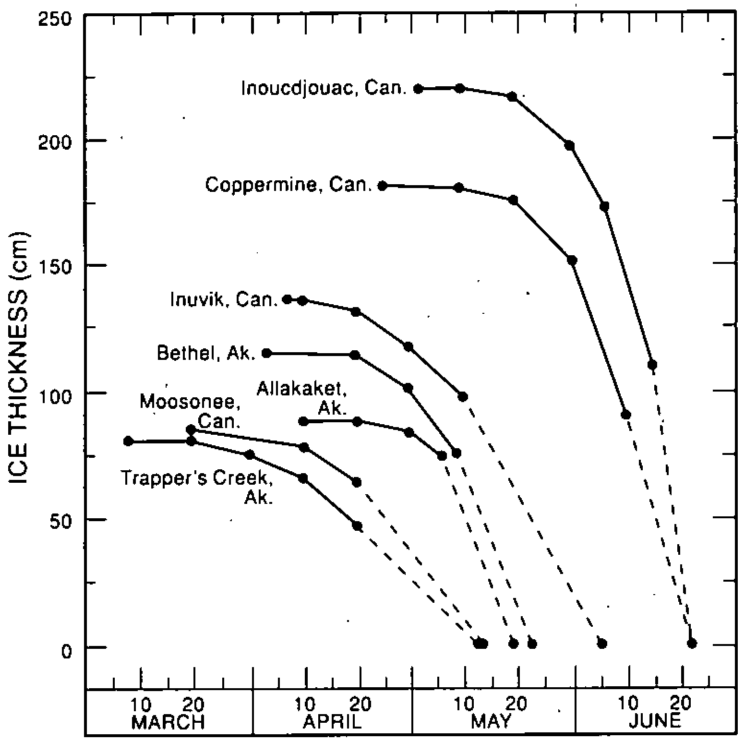


Figure 2.24 Average ice decay curves for seven northern rivers (from Bilello, 1980)

With rising spring discharge, both flow depth and velocity increase. At the same time, the ice ripples noted above begin to develop at the base of the ice cover, thereby increasing the roughness and the rate of heat transfer. Hence, even with near-zero water temperatures (and while mean air temperature can still be negative; see Bilello 1980), ϕ_K can be significant if effective heat transfer can be generated through high flow velocities and/or a rough ice bottom.

Under "normal" conditions (i.e., where the channel is totally ice-covered), thinning is usually < 1 cm/d; however, much higher rates have been recorded in cases of significant thermal input from tributaries or where upstream heating can occur in upstream open-water sections.

Higher water temperatures are related to the amount of incoming short-wave radiation and the expanse of open-water area. Calkins (1984), for example, found rates as high as 8 cm/d for a relatively shallow river during the spring melt period in Vermont. Furthermore, Prowse and Marsh (1989), found that an input flow temperature of approximately 1°C, moving beneath a 4 km reach of a thick ice jam at a velocity in the order of 0.4 m/s, produced an equivalent thinning rate of over 1 m/d.

The combination of terms controlling subsurface thinning has special significance for rivers prone to large variations in winter discharge, such as those found under regulated regimes. Releases of warm water from reservoirs, increases in velocity, and changes in the stage and velocity distribution can all lead to rapid thinning of an ice cover and the formation of a significant surface-transport hazard.

2.8.5 Open Water Sections and Polynyi

Under spring melt conditions, combinations of hydraulic and thermal conditions usually lead to the formation of open water sections prior to the actual break-up of the cover. Such conditions, however, can also exist throughout the winter period, leading to the retention of open-water sections, or polynyi. Polynyi formed by warm groundwater from streambed inflows or laterally from hillslopes, or by friction heat through turbulent rapids are common examples.

From a thermodynamics perspective, such holes can be especially significant since they are, essentially, a window to an otherwise sealed system. For example, open rapids offer ideal sites for significant frazil-ice growth, hence the subsequent downstream formation of hanging dams. Free convection over any form of open-water sections also favours the development of fog and ice-fog under cold atmospheric conditions. Such fog can modify the radiation regime and convective fluxes over the open-water section (e.g., Seagraves, 1981; Wendler, 1969). Where the open-water areas are extensive, such as downstream of hydro-electric plants, the fog can become a natural hazard because of its effect on visibility.

Polynyi have long been recognized sites of intense biological activity in marine environments and, more recently, in northern rivers, particularly as they relate to fisheries habitat. Mapping of open water areas has typically been conducted in conjunction with fisheries overwintering studies (Bryan, 1973; Northern Natural Resource Services, 1979; Blachut, 1988) due to the apparent correlation between overwintering fish habitat and reduced ice conditions and/or the typically warmer groundwater flow. Notably, these areas have been reported to be the only known overwintering and/or spawning areas for a number of fish species (Bryan, 1973). Given that they are the only locations at which air-atmosphere exchanges can occur, they have broader environmental significance, such as in re-oxygenation of water.

2.8.6 Meltwater Release

As described in Section 2.8.4, pre-break-up thinning of the cover is usually less than approximately 1 cm/d and, as a result, only small additions to flow can be expected (e.g., Williams, 1971). Much greater quantities of meltwater may be rapidly released to the spring flow at break-up as a result of high leading-edge water temperatures and rates of melt, such as described in Sections 2.8.3 and 2.8.4. During the rapid movement of a break-up front, additions of flow from meltwater are likely to be overshadowed by contributions of water released from backwater storage (Section 2.3.2), but once an ice cover has jammed, meltwater release becomes most significant. For example, in a study of the thermal decay of a large ice jam, Prowse (1990b) found that approximately 30 x 10^6 m^3 of water was melted from the jam in a three-day period. This approximate 100 m^3/s addition to flow augmented the spring discharge of the river by about 5%. The greatest augmentation of spring flow can be expected during dynamic break-ups that lead to both large ice jam formations and high water temperatures.

2.8.7 Underwater Spectral Radiation

For the purposes of calculating under-ice heat budgets, short-wave radiation was treated as a single, uniform flux. Snow and ice are, however, spectrally selective (e.g, Grenfell and Maykut, 1977; Warren, 1982). Although the magnitude of spectral filtering is small relative to the overall reduction in light quantity, at very low levels of radiation, spectral selectivity becomes increasingly important for biological activity. This is particularly true for the 450- and 660-nm wavelengths that favour photosynthesis. Figure 2.25, for example, illustrates the spectral dependence of radiation absorption in granular ice.

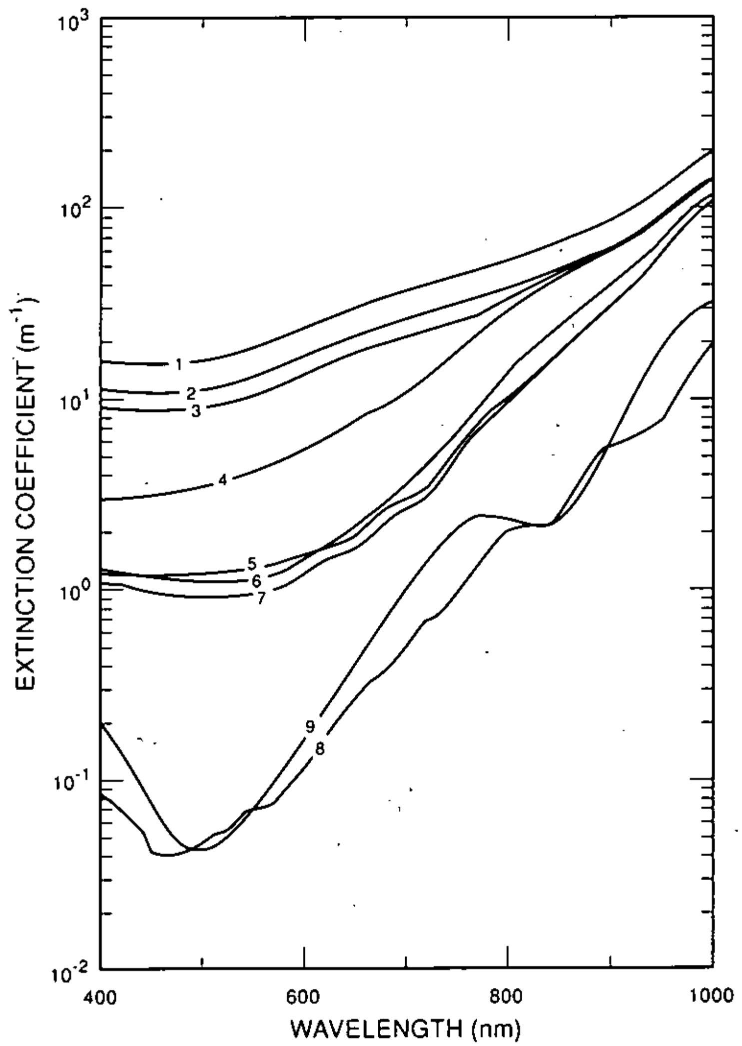


Figure 2.25 Spectral extinction coefficients for nine medium types used in the sea ice radiative transfer model; 1) dry snow, 2) melting snow, 3) ice colder than the eutectic point, 4) surface scattering layer of white ice, 5) interior portion of white ice, 6) cold blue ice, 7) melting blue ice, 8) bubble-free fresh ice, and 9) clear arctic water (from Perovich, 1989)

Spectral modification to the incoming short-wave spectrum results both from spectral scattering and absorption. Solar angle and surface structure determine the spectral composition of solar radiation penetrating the ice-cover surface. Subsequent scattering and attenuation is dependent largely on internal ice structure (e.g., grain-size, shape and orientation; Grenfell and Perovich, 1981; Maguire, 1975a,b; Warren, 1982).

The size and distribution of bubbles have also been found to be important factors in scattering. The degree of scattering increases with bubble density but the overall effect decreases with wavelength (Mullen and Warren, 1988). Although no work has been conducted on the optical role of melt pores (e.g, Prowse *et al.*, 1990), they should produce similar spectral modifications to these air bubbles. Biologically, however, they may be more significant since they are most prevalent during the spring melt period, a time of significant under-ice biological activity. Implications of ice-modified radiation regimes are further discussed in Chapter 3 and 4.

2.8.8 Climatologic Effects

Macro-scale climatologic effects of floating ice covers are best understood for the case of sea ice, particularly in terms of the regulation of the convective and radiation energy exchanges controlling the Arctic climate (e.g., Barry, 1983; Clark, 1982). Although snow plays a similar role in affecting the terrestrial portions of northern climates, floating ice covers within the terrestrial landscape are essentially surface anomalies and their climatic effects are spatially reduced to that of the meso- or micro-scale. The largest-scale effects are those produced by the ice covers of large lakes.

A climatic regime is often described in terms of the magnitude and variability of the air temperature and humidity. Atmospheric heating can be modified by surface anomalies through changes in latent and sensible heat fluxes, the absorption and reflection of solar radiation, and the emissivity of the surface. The degree to which the atmosphere is influenced depends on the magnitude, scale, timing, location and duration of the anomaly (Walsh *et al.*, 1985).

Although climatic effects of terrestrial water bodies are well documented for open-water conditions (e.g., Nemec, 1973), much less is known about the ice-covered period. Existing information focuses primarily on transition periods, when an ice- and snow-covered water body produces a significantly different energy balance than that of snow-covered land. For example, Rouse *et al.* (1989) showed that the presence of ice greatly influences the intertidal zone of Hudson Bay. Ambient air temperatures were up to 8°C colder above the ice zone than over the adjacent snow-free terrestrial area. Onshore winds were as much as 5°C lower than the offshore winds and the vapour-pressure deficits were more than three times smaller. Similar effects occur around large lakes where the near-shore climate is usually characterized by cooler springs (e.g., Quinn *et al.*, 1980).

The lateral dimensions of large rivers can approach those of small reservoirs and the scale of their climatic impacts are likely to be comparable. The largest impacts are expected for the delta regions of large river systems in which the network of tributary channels and interconnected ponds and lakes blanket a vast area of the landscape. The Mackenzie River Delta, for example, develops from northward-flowing rivers (Mackenzie, Arctic Red and Peel Rivers) with channel widths of less than 1 km to a maze of channel systems over 60 km wide and occupying an area of approximately 6500 km^2.

The conversion of the delta environment during the break-up process has been identified as a critical period in the thermal regime of the regional meso-climate. Two processes control the thermal significance of break-up: (a) the influx of warm river water that accompanies the break-up front, and (b) changes to the surface radiation regime. The magnitude of the hydrothermal flux depends on discharge and leading-edge water temperatures, as previously described in Section 2.8.3-4. Based on comparable data (MacKay and Mackay, 1974), it has been calculated that the total hydrothermal flux to the delta increases from near zero in April to 235 x 10^9 MJ in May (Findlay, 1981), the latter value comparable to that delivered by net radiation for the same month. Arrival of the break-up front leads to rapid clearing of snow and ice within the main flow-channels, and flooding of delta lakes and lowland areas by sediment-laden water. Both processes lead to a dramatic reduction in surface albedo, especially where flooding inundates significant areas of the delta landscape. Under hydrometeorological conditions typical of a dynamic break-up (i.e., a rapid event with little prior deterioration of snow and ice), Hirst (1984) notes that peak inundation of the delta surface varied from 90 to 100%. Under such conditions, local temperatures within the delta rise rapidly by as much as 5°C (Hirst, 1984). By contrast, a thermal break-up (i.e., a more gradual event with extensive deterioration of the ice and snow cover prior to break-up) produced only 30 to 60% flooding across the southern and central delta (Hirst, 1984). Under such conditions, local delta temperatures were affected relatively little. Hirst (1984) further noted that although the differences were generally small, the biological significance could be high. Break-up normally occurs when the mean daily air temperature is near 0°C. Slight increases in radiative warming could raise air temperatures above a critical value that permits budding and plant growth. Gill (1974), for example, notes that bud-burst of feltleaf willow in the delta occurs two to three weeks earlier than in nearby tundra locations not affected by the meso-scale climatic effects of break-up.

As the scale of the surface anomaly decreases, so does the climatic effect. Hence, ice covers on most small streams and rivers are likely to produce a limited local effect, perhaps significantly affecting only the riparian zone. An exception would be where large-scale icings dominate large portions of river valleys. Their persistence throughout the summer and effects on local meteorologic conditions can produce a cooler local climate, known to be favoured by some biota. Large mammals are also known to favour icings during the winter months. Gill and Kershaw (1979), for example, found that moose and wolves concentrate along icings which act as snow-free corridors for food access.

3

EFFECTS OF RIVER ICE ON CHEMICAL PROCESSES

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3.1 INTRODUCTION

Although the physical effects of river ice have received considerable study, as evidenced by the various detailed reviews in Chapter 2, the chemical aspects of river ice have not been so well documented. The overall effect on water quality during the ice-covered period is especially important because this is also when the riverine ecoystem becomes stressed by a combination of environmental factors related to low winter flow, minimal solar radiation, and prolonged sub-zero temperatures.

This chapter first provides some background about the basic concepts of river chemistry and some unique characteristics of ice. Focus is then placed on ion exsolution and elution, and seasonality of ionic concentrations. Discussion is also included about the importance of seasonal flow variations and the effects of snowmelt, particularly as they relate to different periods of the ice season, i.e., formation, growth and break-up. Special emphasis is then given the role of ice cover in affecting dissolved oxygen levels and chemical reaction rates. Finally, the implications of these findings are discussed and future research needs identified.

3.2 WATER CHEMISTRY

There are eight major, naturally occurring ions in freshwater systems: the cations consisting of sodium (Na^+), potassium (K^+), magnesium (Mg^{2+}) and calcium (Ca^{2+}): the anions consisting of carbonate and bicarbonate (CO_3^{2-} and HCO_3^-), sulphate (SO_4^{2-}), chloride (Cl^-) and phosphate (PO_4^{3-}). Other important river water components are oxygen, nitrogen (N_2, NH_4^+ and NO_3^-), hydrogen (H^+), silica, and the large variety of dissolved organic substances resulting from synthesis and decomposition by aquatic organisms.

The concentrations of all of these solutes are remarkably variable, even in undisturbed aquatic systems. The physical geography and geology of the region, river and bank flora and fauna, and temperature and light are all known to have an effect on water chemistry. The origin of the water in a river, be it mainly from groundwater, surface run-off or precipitation, also plays a role in determining the river hydrochemistry. The ionic content of groundwater-fed systems depends on the origin of the water and the geology of the ground through which it travels. For example, bedrock water may be more saline than groundwater from glacial drift due to the exchange of calcium and magnesium for sodium as the water flows through the rock.

Dissolved oxygen and organic and inorganic nitrogen and phosphorus concentrations play a role in predicting the type and level of activity of aquatic organisms and are, therefore, important parameters in describing water quality. The concentrations of these and other components of natural waters are studied to determine their effects on aquatic life and on the suitability of water for different uses. The following sections describe both how water quality affects the chemical nature of ice and its meltwater, and how river ice controls specific aspects of water quality.

3.3 ICE FORMATION AND ION EXSOLUTION

As discussed in Chapter 2, the physical processes involved in ice formation are governed by the physical characteristics of the river, the flow regime, and climate of the region. Chemical processes, however, can also affect ice formation and related water chemistry. For example, ice has been noted to form even from humic foam in the case of heavily polluted water or water with a high quantity of organic matter (Ettema *et al.*, 1989). The foam is usually created by fast-flowing water agitating the matter in the water and causing it to bubble. This occurs on streams where there is a lot of foam floating on the surface, which freezes during low temperatures to form "foam pans"; the ice formed this way somewhat resembles white ice and is high in ionic content. Such ice might even reduce light for photosynthesis and cause the levels of dissolved oxygen to drop earlier in the season than when black ice forms on the stream (see Section 3.9 for overall discussion of ice-oxygen effects).

Although humic-foam ice is an extreme case, the presence of even small amounts of solutes in river water can affect ice formation and related water quality. In general, the presence of most solutes in the water weakens the hydrogen bonding between water molecules, increases intermolecular distances, and lowers the freezing point of the solution. Ionic solutes impose considerable strain on the lattice structure and are usually expelled. The ions partition themselves between the water and the ice according to a distribution coefficient K_d which is equal to the concentration of the solute in the ice divided by its concentration in the underlying waters (Hobbs, 1974).

As ice grows it tends to increase the solute concentration of the underlying water. For example, growing sea ice is known to cause the concentration of brine in the water directly underlying it to increase up to almost three times its original level (Wakatsuchi and Ono, 1983). The total ionic strength of fresh water is lower than that of salt water, but similar ion exsolution is presumed to occur. Field studies of arctic lakes have shown that ice excludes over 90% of the dissolved oxygen as well as a large percentage of the other substances dissolved in the ice-forming water (Welch, 1974). Because of constant mixing and throughflow, the effects may be less apparent in rivers, but all waters affected by this process should become more concentrated in ions, gases and organic matter. This freeze-out process is known as cryoconcentration; the exsolution of water ions results from this process.

Experiments both *in situ* (Catalan, 1989) and *in vitro* (Brimblecombe *et al.*, 1987) have shown that certain substances are preferentially exsolved during ice formation, whereas others are preferentially eluted during ice melt. Dissolved gases and un-ionized particles are virtually totally excluded from growing ice. In experiments, it was found that the order of ion retention in ice is: $Cl^- >> Ca^+ > HCO_3^- > Mg^{2+} > Na^+$. The Cl^- is high in retention, since it aids in the formation of ice. The elution sequence seems to be $SO_4^{2-} > NO_3^- > Cl^-$ and $Mg^{2+} = K^+ > Na^+$. Moreover, it was determined that the elution of anions was not proportional to the elution of their respective counter-ions.

Solutes such as phosphates show no preferential partitioning between ice and water, the ice containing roughly the same phosphate concentration as the water. Certain ions such as chloride, for example, can aid in the stabilization of ice crystal lattices and are therefore incorporated into the growing lattice structure. Alkali metals, on the other hand, are crystal structure breakers that tend to be expelled from growing ice. Thus, the ion content of the ice bears little relationship to the total ionic content of the river water.

Efficiently eluted ions are highly concentrated in the initial meltwater and become rapidly depleted, whereas other less easily removed ions dominate the more dilute meltwater. The measurement of specific expulsion in the field is very difficult due to the simultaneous presence of inflows, outflows, and precipitation, and also to the fact that preferential elution does not take place until well into the melt.

Preferential freeze-out is thought to play a role in the annual retention of solutes by the river system. Substances such as phosphorus and NO_3^-, which are very inefficiently exsolved (as little as 30% removal), are not well retained by a river when the ice-bound chemicals are carried downstream and lost during flood flows in the spring.

The efficiency of exsolution is highly dependent on the rate of ice formation. Rapid growth does not permit the vertical transport of denser solute-rich water to the waters below. Rapidly-formed ice contains more impurities than ice grown more slowly. The relative inefficiency of ion freeze-out can be indicated by the presence of bubbles in the black ice. Preferential elution is also an indication of inefficient ion exsolution. When the growth rate of black ice decreases, freeze-out becomes less ion specific.

3.4 SEASONALITY OF IONIC CONCENTRATIONS

The concentrations of most ions in the water column increase throughout the winter, with most reaching their maximum levels under early spring ice (Figure 3.1). This behaviour is exhibited by parameters such as calcium, magnesium, silica, total alkalinity, and total conductivity. This pattern of high alkalinity is shown in the inflows of a number of small Ontario lakes (Molot *et al.*, 1989). The peak was considered to be caused by the early melting of ice and snow during which most of the ions are released as they raise the melting point (see subsequent discussion in Section 3.8 regarding the importance of snowmelt). There are several exceptions to this observed variability. The amount of dissolved oxygen, for example, typically continues to decrease as winter progresses, only to recover slightly just before the end of the winter when the ice cover once again allows light penetration, photosynthesis resumes, and extra oxygen is carried in by the fresh meltwater.

In lake-fed systems, the amount of iron and trace minerals in the water column appears to decline over the winter, presumably as a result of reactions and adsorption at the surface of the sediment. Because metal contents are generally related to sediment, and because erosion processes contribute, therefore, to their in-stream concentrations lower metal concentrations can be expected in winter than in summer (Whitfield and Whitley, 1986).

Sulphate and potassium exhibit a more variable pattern. Potassium seems to show a complex hysteretic behaviour with respect to discharge. It continues to increase in concentration even after spring melt. This is thought to be due to the predominance of the ion in surface snowmelt run-off (Schwartz and Milne-Holme, 1982). The changes in sulphate content, however, were found to be groundwater-dependent in the Yukon River basin study. Post-freshet sulphate concentrations increase with the increasing dominance of groundwater (Whitfield and Whitley, 1986).

The levels of other ions decline to their yearly minimum soon after the break-up of the ice cover due to the dilution effects of high spring flow. This cycle seems to be the major controlling factor in temporal ion fluctuations. Nutrient variability seems more likely to be mediated by biological activities although break-up flooding may play a role in the seasonal input of flooding as noted in Section 2.7.3.

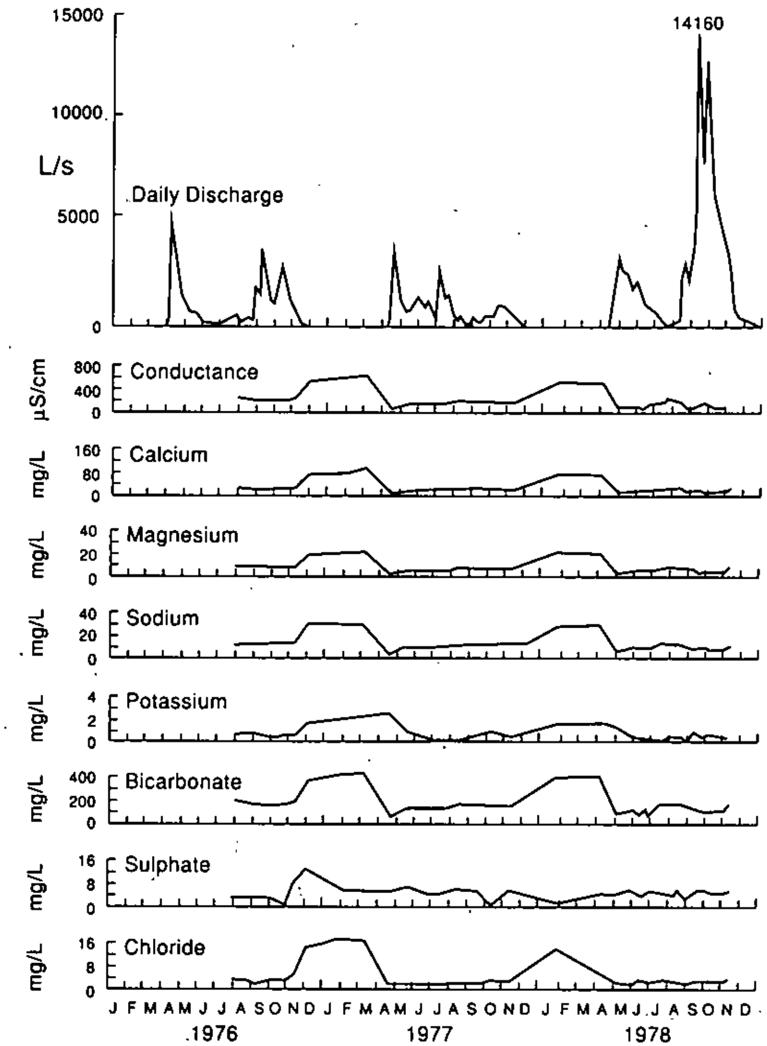

Figure 3.1 Geochemical and discharge hydrographs for Hartley Creek (Schwartz and Milne-Holme, 1981)

3.5 INFLUENCE OF DISCHARGE ON WATER QUALITY

The chemical composition of both the ice and underlying water depends on the original water quality during ice formation and subsequent growth. Water quality, however, varies temporally and has been related to seasonal variations in discharge. For example, in the Yukon River basin (Whitfield and Whitley, 1986), four groups of parameters have been suggested: those such as suspended solids, metallic ions and total phosphorus having a positive dependence on discharge and a clockwise hysteresis with respect to flow; those such as alkalinity, total hardness, specific conductivity, silicate, sulphate and total inorganic carbon having a negative relationship with discharge and a counterclockwise hysteresis with respect to flow; those such as chloride, fluoride, potassium and total organic carbon exhibiting patterns not linked to discharge; and those such as pH and certain nitrogen compounds whose levels are not related to discharge and exhibit a more random variability. As described in Sections 2.2 and 2.4., ice formation and growth occur while discharge is declining, possibly even to its annual low flow point. It is the water quality of this period that governs the ice-water chemistry.

During the spring melt period, which includes both ablation of the winter snowpack and break-up of the ice cover, river water quality is often signficiantly influenced by the chemistry of the melting snowpack (see further discussion in Section 3.9). The relationship between snowmelt and water quality, however, can be complex. For example, in a study of suspended solids released during snowmelt on Gleen Creek, Alaska (Chacho, 1990), it was discovered that the major release of these solids occurred well after the initial snowmelt release. The main portion of solids were not released until the creek was flushing out the last remaining ice and other frozen debris. The change was quite dramatic, from a standard level of about 50 mg/L during the rest of the season to a sudden peak of 1337 mg/L in a day, at which time flow was only at 33% of peak flow. If this pattern is followed elsewhere and not just coincidental to this stream, then this sudden imbalance of solids content could have a harmful effect at the end of winter on water quality and biological life in a stream.

Water quality is also closely tied to sediment concentration in a river. In a systematic study of relations between sediment concentration, C, and river discharge, Q, for single hydrologic events (a brief flood or a relatively lengthy snowmelt run-off), Williams (1989) defined five major common classes of C-Q relations in terms of mode, spread, and skewness following plotting of temporal graphs for C and Q. They include single-valued line, clockwise loop, counterclockwise loop, single line plus a loop, and figure eight. Many factors can affect C-Q relations, such as hydrological characteristics of the event, geomorphological features of the river, amounts and sizes of existing bed- and suspended- materials in the river, and land-use activities along the river-channel banks. When these factors interrelate with other regional and local conditions, the task of predicting the type and magnitude of C-Q relations for a particular site and occasion involves many variables and becomes very complex. Notably, however, such relations have not specifically focused on the period, the major hydrologic event for many northern rivers. As noted in Section 2.7.2, measurements of sediment concentration are extremely rare during river ice break-up. Data that are available, however, suggest that suspended sediment concentrations may exceed those that occur under open-water conditions with equivalent discharge. Two major reasons to explain the enhanced suspended sediment concentation are the exposure of additional bank area to potential erosion by the elevated water levels and accelerated erosion because of intensive ice-bank/bed interaction and high velcities (Prowse, 1993).

Although discharge is only one of several variables governing water composition, these observations may aid in understanding the ionic relationships in the river and in determining the relative importance of ion exsolution and other mechanisms affecting water chemistry. Thus, the typically low winter flow of many rivers can influence the composition of the ice formed.

3.6 ROLE OF CLIMATE

In systems where winter flow is not primarily composed of groundwater inflow, climate would be a primary factor influencing water and ice composition. After the formation of the initial ice cover, the chemical composition of both the water and the ice can continue to change throughout the winter. As ice progressively thickens, there is a simultaneous increase in the water's solute concentration. The change in the water's ionic strength is proportional to the volume of the river water that freezes. In a shallow river, the ice-to-water ratio can be very high, and the underlying water can become very concentrated in exsolved substances.

Other winter events such as snowfall, thaws and thermal ice cracking can affect the continued growth and composition of both the water and the ice. Snow cover will drastically reduce the amount of light that penetrates the ice surface (see discussion of the filtering of short-wave radiation in Section 2.8.2). The resulting decrease in algal photosynthesis is associated with an increase in bacterial decomposition activity and with a decline in the amount of dissolved oxygen, as discussed in Section 3.9. The cessation of photosynthetic activity is almost immediate after the accumulation of any significant amount of snow cover (Prowse and Stephenson, 1986).

Ice cracks may be caused by a number of forces, including the pressure created by significant flow fluctuation under ice cover or by a heavy snowfall, particularly one occurring after a mid-winter thaw (see previous discussion of snow-ice or white-ice formation in Section 2.1.4). Positive hydrostatic pressure in the water column will then force ion-rich water from underneath the ice to spread over the ice surface. Upon return to freezing temperatures, the resulting slush forms a layer of rather highly conductive ice on top of the existing ice cover. Over the length of the winter, successive slushing events can help to create layers of ice varying in chemical composition intercalated with a slushy solute-rich region (Jones and Ouellet, 1983).

This zone of white ice is the result of a complex interaction of meteorological phenomena (snow, rain, solar radiation), meltwater percolation, and the diurnal freeze-thaw cycle. The outcome of cracking and slushing events is a highly heterogeneous layer of snow, ice and slush. Black ice, therefore, acts as a very non-selective permeable layer that even allows microorganisms to be transported upwards. Their presence can be demonstrated by the detection in the slush of measurable quantities of adenosine triphosphate, a product of microbial metabolism.

At the end of the winter, it is also possible for negative hydrostatic pressure to occur, forcing the slush through cracks into the under-ice water column. Solutes contained in the slushy white ice zone can, under such conditions, be returned to the water column.

The net effect of these various ions and water movements, being a function of climate, will be different for each river and for each year. In areas characterized by very cold temperatures and light snowfall, little white ice will be formed and the total ice cover will be very low in ionic content. This would have several effects. Depending on the amount of sunlight present during the period of ice cover, light will continue to penetrate to the water column, allowing photosynthesis to occur.

The depletion of oxygen might not be as severe under such circumstances, but the ionic concentration of the water will probably not be relieved by transport due to the lack of slushing, or by molecular diffusion through the crystalline lattice. Therefore, increases in ice thickness cause an increase in ionic strength wherease factors, such as precipitation and adsorption of solutes onto the river bed, cause a decrease. Groundwater inflow can cause either an increase or decrease, depending on the ionic content of the surrounding soil. Figure 3.2 illustrates the change in oxygen content in rivers during winter.

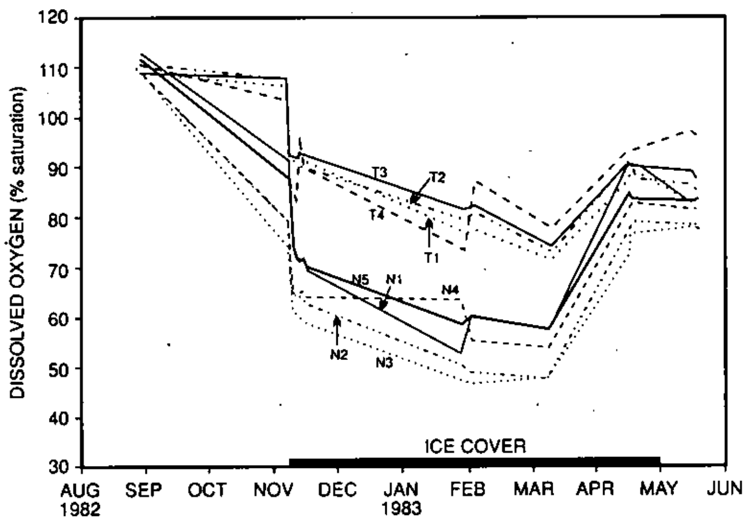

Figure 3.2 Dissolved-oxygen concentrations in the Takhini and Nordenskiold rivers from August 1982 to May 1983 (Whitfield and McNaughton, 1986). Period of ice cover is indicated by the heavy base line. T1 to T4 and N1 to N5 are sampling points on the river.

In regions where slushing predominates as an ice-forming mechanism, the total conductivity of the ice could be much higher. The total amount of under-ice oxygen could be greatly reduced as oxygen-rich water leaves the water column to form the new white ice. The presence of the white ice will effectively inhibit photosynthetic activity,

compounding the oxygen-depleting effects of the slushing. If temperatures permit, ions will also migrate through the black ice, reducing the water's ionic content.

The rate of spring melt will also temporarily affect the chemistry of the water. A slower melt from the top down when multi-layered white ice is present will result in a higher concentration injection into the water column. A rapid melt when both black and white ice melt at the same time will not show such an ion injection. The contribution of ions to the water or their dilution during the spring is highly dependent on the proportions of the black ice, white ice, slush and snow that make up the winter cover.

3.7 OTHER TRANSPORT MECHANISMS

Other ion transport methods are also presumed to occur in rivers. Molecular diffusion due to the concentration gradient caused by preferential ion exsolution is thought to take place in ice. As well, it has been observed, through the use of fluorescent dyes, that black ice contains very small interstitial canals whose diameters depend on the ambient temperature. During mild periods, the ice becomes permeable to ions from the water column. With time, the black ice could contain increasing amounts of impurities. The size of these interstitial canals would limit this incorporation to ions and other small molecules.

Migration can also occur through the layer of snow blanketing the ice surface. Impurities of atmospheric origin can accumulate near the ice surface and be assimilated later either by migration or by slushing. Atmospheric water, being very different in composition from river water, can introduce to the ice a variety of foreign contaminants that can be incorporated into the water during the spring melt.

The incorporation of atmospheric pollution could be especially devastating to river biota if the precipitation were acidic (Gunn and Keller, 1985). The fact that the first 10% of snowmelt can contain up to 50% of the solute concentration of the winter snow cover can, in the case of an early or midwinter thaw, cause a dramatic temporary decrease in water column pH (Baron and Bricker, 1987). In many cases, the snowmelt precedes icemelt by some time, causing the late winter or early spring jump in ionic content of stream water. For example, in one case, a brief thaw in February caused a 40% decline in the Ca^{2+} concentration in the snowcover (Stottlemyer, 1987). Thaws subsequent to the first major melt would contain a lower concentration of contaminants and therefore would be less harmful to the life of river organisms.

Given the concomitant melt of snowcover and river ice during the spring break-up, and the potentially high influence of snowmelt chemistry during this period, it is useful next to review briefly the effects of snow on river water quality.

3.8 EFFECTS OF SNOWMELT ON WATER QUALITY

Numerous studies have been done on the effects of snowpack and spring melt on northern water bodies. In general, the chemistry of stream and lake waters can be influenced by the input of concentrated meltwaters and the amount of rainfall on the snow cover. The decrease in pH in the water column and its effects have been noted, for example, in several investigations of northern-Ontario Canadian lakes (Gunn and Keller, 1986; Semkins and Jeffries, 1986), and at an upland catchment in Scotland (Abrahàms *et al.*, 1989). Aluminum leached from the soil during snowmelt has also been associated with fish toxicity. This toxic effect appeared to increase with low pH and low calcium ion levels, both of which can occur in the river at the time of spring melt. The fact that these events occur at the same time as the embryonic development of fish in the river has caused concern about species survival and the potential for mutation.

In areas where severe acid precipitation occurs, the levels of ions added to the river could be considerable, even devastating if there is just enough snowmelt to release this sudden input of ions into the river water. Roads along rivers can also have a negative effect on river chemistry. Material containing road salt is often ploughed to the side of the road after a snowfall where it can melt directly into a river in early spring, causing a large input of ions. In the case of a smaller stream, this could have a deleterious effect on aquatic life. In one study, measurements at a test site near a road created a distortion in all of the average measurements because of this sudden input of ions from road salt (Molot *et al.*, 1989).

Surges in the concentrations of nitrates and sulphates have also been noted, although the implications of this observation are not yet well defined. It has been shown by Abrahams *et al.* (1989), however,

that a preferential elution similar to that in icemelt is at least partially responsible for these peaks. More specifically, Lockerbie (1988) studied the impact of winter snowpack and its melt on the chemical characteristics of the receiving water in the upper Mersey River in southwest Nova Scotia, during the three winters from 1984 to 1987.

In general, temporal evolution of snowpack solute loading over the winter period is governed by two main processes: (a) decreased snowpack ion loading results from either above-freezing temperatures and/or rain or mixed precipitation inputs that have a lower ionic loading than the existing snowpack and cause an overall decrease ionic loading of the pack through a dilution effect, and (b) increased snowpack ionic loading results from new snow inputs having a higher ionic loading relative to the existing snowpack, or, to a lesser extent, from sublimation of the existing pack. Spring melt of the snowpack in the basin causes a short-term acidification of river water through two processes: a dilution of base cations, and a disproportionate influx of acid anions. Episodic "spring acidification" resulting from relatively high concentrations of sulphate in the river water is also augmented by sulphate and nitrate loadings from the melting snowpack.

Although most winters in southwestern Nova Scotia are relatively mild, usually characterized by several freeze-thaw cycles during the winter months, the upper Mersey River basin contains several lakes that typically freeze during the winter months. However, studies to date have not been designed to differentiate between the effects of snow/ice melt directly on the surface of the river and on the overall basin.

The release of ions in the snow itself during melt closely resembles that of ice. The general order of elution is $K^+ > Ca^{++} > Mg^+ > Na^+$. For cations, sulphates and nitrates are eluded before chloride (Brimblecombe *et al.*, 1987). The difference between the ice and the snow is that the snow holds a larger quantity of ions and does not preferentially elude them until spring melt.

3.9 CHANGES IN DISSOLVED OXYGEN LEVELS

The oxygen content of water is of foremost interest when dealing with water quality. The dissolved oxygen criterion established by biologists as the level necessary to maintain aquatic life in freshwater systems is approximately 7 mg/L. Well-aerated waters can be saturated with dissolved oxygen. Oxygen supersaturation can even occur in natural waters when photosynthesis rates are higher than the rate of oxygen usage. This saturation is temperature-dependent but is usually sufficient to meet the criterion.

In southern rivers such as the Mississippi and those in southern Ontario, dissolved oxygen concentrations have been known to decrease to low levels during the summer, especially at night when photosynthesis does not occur. Although waters in colder regions do exhibit a similar summertime drop, a more dramatic decrease under winter ice cover has been observed. This event is thought to be the major cause of the winterkill of fish in ice-covered lakes (see extensive discussion of under-ice oxygen/biological examples in Chapter 4).

Varying types of oxygen depletion conditions have been observed during the study of lakes in the High Arctic (Welch, 1974) which are ice-covered for a large portion of the year and can often be characterized by low inflow and groundwater discharge, and the study of other ice-covered lakes in more temperate zones (Babin and Prepas, 1985). In lakes with low respiration rates, the rate of oxygen freeze-out can be the same as or slightly higher than the rate of lake oxygen use. In lakes with high metabolic rates, the concentration of oxygen can decrease over the course of the winter. Nearly anaerobic zones within 0.5 m of the bottom have been reported (Barica and Mathias, 1979).

Severe dissolved oxygen depletion has been found also in Alaskan (Schallock and Lotspeich, 1974) and Yukon rivers (Schreier *et al.*, 1980; Whitfield and McNaughton, 1986). A number of rivers, varying in location, size, drainage area and flow discharge, have exhibited some degree of winter oxygen decrease. Certain patterns have been noted: one concerns temporal changes, and the other, spatial variations. Because of a lack of reaeration, the levels of dissolved oxygen in the stream drop. Unlike lakes where oxygen levels vary primarily with depth, the main changes in the river are due to distance of travel downstream as oxygen is used up with distance and none added until another input to the river. An example of this is shown for the Yukon River in Figure 3.3 (Schallock and Lotspeich, 1974).

The seasonality of dissolved oxygen concentrations is characterized by depression from near-saturation conditions just prior to freeze-up, to annual low levels under ice cover in the late winter before the resumption of photosynthetic activity. As photosynthesis rates increase under the ice and during ice break-up, the production of oxygen increases, allowing water-column oxygen concentrations to return to

high fall values (Figure 3.4). Also, the inverse relationship between water temperature and dissolved oxygen concentration found in temperate zones is only evident during the short summer period in arctic and subarctic regions.

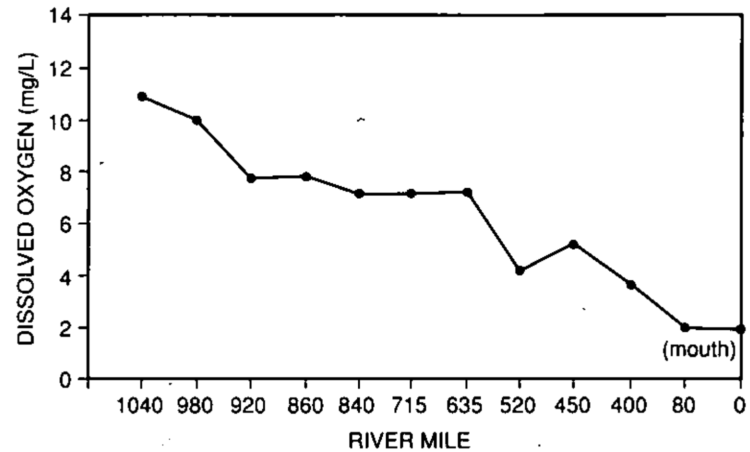

Figure 3.3 Winter dissolved oxygen profile for the Yukon River (1971 data) (Shallock and Lotspeich, 1974)

Spatial variations within one river indicate that headwater oxygen content is normally greater than the downstream oxygen content. Geographically, widely separate rivers can have similar oxygen-depletion patterns, whereas very close rivers may exhibit vastly different depletion levels. Any natural water system that freezes can exhibit some degree of oxygen depression, although the presence of regions of open water can allow reaeration to occur. Even in the High Arctic, such open water has been observed.

The wintertime depletion of river oxygen is the result of the interaction between a great number of chemical, physical and biological factors. Reduced under-ice oxygen levels can be attributed to lower reaeration due to ice cover, lower photosynthesis rates in response to low temperature, reduced light (Section 2.8.7) and lower availability of nutrients, and the reduction in flow available for dilution associated with winter conditions.

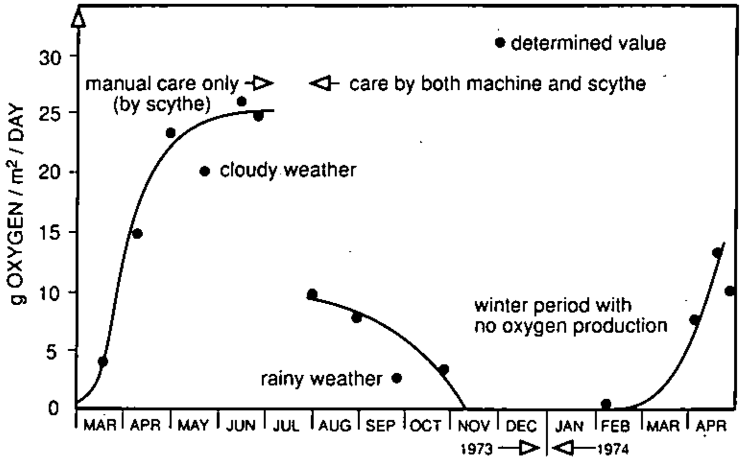

Figure 3.4 Seasonal variations of the total daily gross production, for a reach of the Havelse stream (Simonsen and Harremoes, 1977)

Nutrient level can also influence the rate of oxygen depletion under ice conditions (Choulik, 1988). In arctic regions, eutrophic lakes exhibit more rapid and more severe depletion than do oligotrophic ones. The higher demand for oxygen in nutrient-rich waters is normally expected, as eutrophication can be associated with an increase in microbial decomposition activity that requires a great deal of oxygen.

Although the highest oxygen saturation levels occur at the lowest temperature, it does not follow that water at these temperatures has the greatest affinity for oxygen. As the temperature decreases, the reaeration coefficient also decreases (Eheart and Park, 1989). This effect could multiply the depletion of oxygen as ice cover forms because there is a smaller area for the water to be reaerated.

Similarly, oxygen depression in rivers subject to long periods of ice cover is the most notable result of municipal waste loading (Bouthillier and Simpson, 1971). It has also been noted that the decay of faecal coliform bacteria, an indicator of waste load decay, is almost negligible under ice cover (Putz *et al.*, 1984). This raises concerns about the

possible health effects of wastewater discharge in cold regions. In the sparsely populated arctic and subarctic regions, the current waste loading is very low but still has the potential to influence dissolved oxygen content of rivers.

Another cause of winter oxygen depletion is the oxidation of carbonaceous matter, organic and inorganic nitrogenous matter, and inorganic materials in the water column and sediment, the relative amount of these substances varying from river to river and from year to year. For example, oxygen depletion under ice cover has been observed in power dam headponds on the Saint John River, New Brunswick below point source inputs from pulp mills and food processing plants. This is also the major focus of a major research project on the Athabasca River of Alberta.

The respiration of aquatic and benthic flora and fauna can further reduce the under-ice oxygen level. Due to reduced light levels under ice cover, even the oxygen-producing photosynthetic bacteria must use rather than produce oxygen to survive.

The benthic organisms capable of nitrification — that is, the conversion of ammonia nitrogen to nitrite and finally to nitrate — are the principal cause of sediment oxygen demand. Under nearly anoxic conditions, these organisms can also release NH_3, CO_2 and H_2S into the water column, which can have detrimental effects on the life of fish and other aquatic beings.

3.10 CHANGES IN REACTION RATES

Each chemical reaction is characterized by a reaction rate. This reaction rate can be affected by the concentration of the reactants and products, the presence of catalysts, and the reaction conditions. Temperature and pH can play an important role in determining the speed of many chemical and biochemical reactions.

Although the dynamics of many of the reactions occurring in natural waters have been studied, these studies have, for the most part, been undertaken in temperate water systems. Kinetic constants for some of the most important reactions such as the biochemical oxygen demand (BOD), deoxygenation rate coefficient (K_1), the reaeration rate (K_2), the rate of loss of BOD due to settling (K_3), and the rate of nitrification (K_N) have been expressed as a function of temperature and other factors. Few tests to verify the validity of these constants over the range of naturally occurring temperatures have been performed at very low temperatures.

The temperature of the water under ice cover can be below 0°C due to the presence of solutes that lower the freezing point of the water. Under such conditions, the BOD deoxygenation rate constant, K_1, has been described as a low and constant value in keeping with the assumption of the rate's temperature dependence (Ranjie and Huimin, 1987). The actual decomposition of BOD is very low under ice conditions. This slow decay can lead to the concentration and accumulation of BOD components under ice cover where the rate of addition of BOD materials exceeds the rate of decomposition.

Until recently, it was always assumed that the reaeration rate, K_2, increased with decreasing temperature. This assumption was based on the fact that dissolved oxygen saturation levels increase with decreasing temperature. Using mathematical models (Chao *et al.*, 1987), attempts have been made to show that certain circumstances could cause the reaeration rate to decrease rather than increase with lower temperatures. Should such attempts succeed, it may be possible to explain some puzzling experimental data. Factors associated with low temperatures, such as decreases in the flow rate, the lowering of the diffusivity of oxygen in water as well as decreases in the water surface available for gas transfer from the atmosphere, could possibly cause this type of reversal in the temperature dependence of the reaeration rate. However, it is thought that the classical assumption is probably still valid in most natural systems (Eheart and Park, 1989).

The settling rate of BOD, K_3, is not directly related to water temperature levels. The settling rate is, however, directly related to the flow rate of the river. Because low flow rates are often associated with winter conditions, settling of BOD will increase. This increase might compensate for the increase in BOD concentration due the low flow rate available for dilution (Eheart and Park, 1989).

Sediment oxygen demand (SOD) has been found to be one of the main causes of winter oxygen depletion in lakes (Mathias and Barica, 1980). It was hypothesized that there is enough mixing of water to keep oxygenated water circulating over the sediment in lakes. Since rivers have much more circulation, this effect could actually be increased. SOD is more reactive under warm temperatures, and in the cases of some far northern streams, if a large quantity of organic matter

enters the river in the autumn, the oxygen uptake from it can overcome the oxygen reaeration (Schallock and Lotspeich, 1974) when the temperature warms and ice melts in spring .

Little investigation has been made of the effects of an ice cover on nitrification. One Ontario lake study presented evidence to suggest that significant nitrification does occur under ice cover (Knowles and Lean, 1987). It was also suggested that denitrification can take place when oxygen is severely depleted, allowing the production of gaseous products. Nitrification was proposed as a major factor contributing to the winterkill of fish under ice-cover conditions.

Sediment oxygen demand includes sediment chemical-oxygen demand and sediment biochemical-oxygen demand, both of which have temperature-dependent decay rates. Sediment oxygen demand can contribute greatly to under-ice oxygen depletion. Low temperature studies of the rate of sediment oxygen use have not yet been undertaken, although it is assumed that the temperature dependence of the reactions will still be valid at near-zero temperatures.

3.11 FUTURE INVESTIGATIONS

The enhancement of solute concentration caused by preferential exsolution could potentially have adverse effects on the life in a river. In one prairie-lake study (Mathias and Barica, 1985), nitrogen gas supersaturation due to expulsion of the gas during the formation of lake ice cover has been suspected as a possible cause of winter trout mortality due to gas bubble disease. However, most Canadian rivers subject to long periods of ice cover are rather low in solute content, therefore making the effects of cryoconcentration as yet virtually undetected. Also, it is assumed that native organisms would have developed some adaptation mechanisms to the cyclic variation of nutrients and other water components.

The potentially serious consequences of oxygen depletion and ionic concentration require further study of the prevalence, gravity, and temporal variability of solute concentrations. The increase in oxygen depletion and cryoconcentration that occurs as one proceeds downstream signal that spatial variations together with temporal variations can have far-reaching effects on aquatic life. Especially important to examine is the role of ice-cover duration on a river's ability to recover from such increases and depletions. Ice cover on high-latitude rivers can last from several months in more southern regions to most of the year in arctic zones, making these studies invaluable. Such investigations have yet to be made in many arctic or sub-arctic rivers.

Because most investigations of the effects of ice cover presence, cryoconcentration, and oxygen depletion have been performed on lakes, future field testing should include more extensive study of rivers and streams. Flow certainly would play a role in determining the chemistry of the river system, even in winter. Already, certain differences in oxygen depression behaviour between lakes and rivers have been noted. Lake studies have reported slow and gradual decreases in dissolved oxygen levels over the period of ice cover (Barica and Mathias, 1979). Models have been developed to relate winter oxygen demand to mean-lake depth.

In the limited number of studies of rivers, the measured dissolved oxygen concentrations have appeared to be severely depleted within days of initial ice cover formation (Whitfield and McNaughton, 1986). Until more extensive investigations of a larger number of rivers are made, the reasons for such differences between river- and lake-oxygen depletion behaviour will remain unknown.

To quantify natural under-ice chemical and biochemical reaction rates, experiments to determine these rates must be undertaken in rivers subject to freezing conditions. The validity of current water quality models in rivers frozen for long periods of time must be examined. Quantification of the concentration increases caused by ion exsolution in ice-covered northern rivers is necessary to assess the impact of increased loading of the system.

Although studies of the effects of acid precipitation on aquatic systems have begun, more thorough investigations are required to ascertain the role of ice cover in the temporary storage of acid and its subsequent release upon the decay of the ice cover. The temporary jump in pH caused by such events has been measured, but its impact on river chemistry and aquatic life remains to be analyzed. The significance of ice and snowmelt on the immediate surface of a river compared to that in terrestrial portions of the river basin should also be investigated.

Future research into the causes and effects of cryoconcentration and winter oxygen depletion could result in the revelation of important information needed for water resource planning and management in cold regions. The environmental impact of naturally occurring events, such

as those related to ice formation and break-up, must be examined before any development can begin. In particular, a better understanding of the effects of river ice on chemical processes is necessary in the study of northern water resources.

4

BIOLOGICAL EFFECTS OF RIVER ICE

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4.1 INTRODUCTION

Secondary biological production, often expressed as wet or dry weight (or energy content) per unit area, or per unit time, is a measure of the efficiency of an ecosystem in retaining energy in animal biomass. Production is the sum of the products of instantaneous rates of growth, numbers, and mean weights (or energy contents) of all organisms (Chapman, 1978). In examining the effects of river ice, any conditions that cause changes in growth, mortality, and density can modify a species' production. Theoretically, secondary production should provide a sensitive measure of ecosystem performance. However, few estimates are available for the lotic environment, and these usually apply to a single, conspicuous species.

The preferred habitat of fish is usually characterized by parameters such as temperature, cover, substrate, water depth, and water velocity.

Habitat suitability varies with life stage, species, and river conditions. Such habitat parameters are considerably different between open-water and winter conditions. River ice and river-ice processes have profound effects on habitat parameters and affect winter survival of fish. For example, Walsh and Calkins (1986) listed four factors affecting winter survival of Atlantic salmon (*Salmo salar*): (a) crowding of juveniles, (b) migration to avoid overcrowding, (c) freezing of redds and juvenile fish, and (d) high velocities at ice-cover break-up and ice-jam release (e.g., see Section 2.5.7 regarding scour from ice surges and Section 2.7 for effects of ice on substrate habitat).

For stream fish species, it is rarely possible to estimate numbers precisely at frequent intervals without disturbing normal behaviour. For invertebrates, quantitative sampling is difficult because of contagious, non-random or unknown distributions, penetration of the substrate, drift patterns, movements, and life-cycle considerations. Growth is the parameter that can be most easily measured assuming that unbiased samples can be obtained at intervals.

In assessing the impact of short-term perturbations to the environment, immediate effects may be subsequently masked by compensatory (density-dependent) changes in growth, or mortality, which show no differences over the long term. This is a problem when trying to assess the effects of river ice or, for that matter, of any other disturbance. The sequence of events leading to ice formation are those that local faunas are selected to contend with. All species exhibit patterns of mortality associated with life-cycle stages in synchrony with seasonal events. These are part of their normal population dynamics. All display a degree of resiliency in their population dynamics to cope with fluctuations in the timing and intensity of such events.

There is a general consensus in the literature that winter and certain ice conditions in particular can be detrimental to stream animals, but whether annual measures of secondary production reflect this is less certain. The most catastrophic consequences of river ice tend to be unpredictable in time and space and, hence, are not incorporated into any studies. Almost certainly, for the reasons cited above, no other publications deal directly with the subject of this review.

This chapter examines evidence that ice impacts running-water animal communities in ways that could alter secondary production. Emphasis is given to reports of winter mortality, although clearly the winter growth component may be equally important. Because

vertebrates attract more attention than invertebrates the available information and detail tend to differ. Consequently, the two groups are reviewed separately herein. The possibility that human activity may change ice conditions and the frequency of catastrophic events are also considered. A concluding section examines the importance of ice and identifies the kinds of studies needed to increase our knowledge of the effects of river ice on secondary production.

4.2 INVERTEBRATES

4.2.1 Influence of Stream Environment and Location

Diversity and density of benthic organisms are low in glacier-fed streams but increase in mountain and lowland streams, especially within the tree-line. Springs represent special situations in which temperatures and flows are moderated and more stable than surrounding lotic environments. They support a much denser and more varied fauna and provide important overwintering refugia for benthos (Craig and McCart, 1974; Harper, 1981; Craig, 1989): Composition of the benthos changes with the severity of the winter, and several orders of insects — namely Odonata, Plecoptera, and Ephemeroptera, as well as Simuliidae — are not represented in the High Arctic (Oliver *et al.*, 1964). In tributary streams to Char Lake, Cornwallis Island, which flow from late June to early September, Stocker and Hynes (1976) reported a very limited benthos. It was dominated by five species of Chironomidae and three species of Enchytraeidae. Since chironomid life cycles often extend over one year, larvae or eggs must tolerate temperatures down to -20°C. They may escape the erosion of the spring spate in the still-frozen stream bed.

Arctic and subarctic invertebrate faunas of lotic environments are those most likely to be impacted by river ice. Alpine and particularly antarctic environments are less accessible to normal colonization processes and support restricted faunas (Vincent and Ellis-Evans, 1989). The hydrologic regimes of arctic (MacKay and Loken, 1974) and alpine (Slaymaker, 1974) streams impose special and characteristic constraints on their faunas. They display predictable, marked, seasonal changes in temperature, insolation, ice cover, and run-off. Those invertebrates having no means to withstand or avoid freezing, unable to cope with long periods of low temperatures and low-light intensities under ice, are eliminated. As the frequency and duration of seasonal freezing increases, the biota becomes more specialized and robust (Harper, 1981). However predictable the seasonal cycle, it is subject to variations in timing and magnitude of events, the effects of which may be important for survival. For instance, Irons *et al.* (1989) observed that autumn rainfall could have a profound influence on the availability of unfrozen "refugia" in an Alaskan stream by increasing groundwater flux.

Miller and Stout (1989) compared the macroinvertebrate communities of an arctic and subarctic Alaskan stream over three ice-free seasons. They found similar numbers (112 vs 138) of species in each. Both communities were dominated by chironomids. Most revealing was that within a small area, species composition changed from one sample period to the next. The average seasonal replacement percentage was 37%, with 39% of the original pool surviving in the arctic stream as against 31% and 50% in the subarctic stream. In discussing their results, Miller and Stout (1989) favoured an intermediate disturbance hypothesis as the mechanism responsible for maintaining local diversity. "If prolonged freezing, anchor and frazil ice formation, and summer floods and temperature variation are the major disturbances causing local extinctions, then the arctic species array must be disturbed often and selected for high dispersal capability". The success of chironomids in these systems appears related to their ecophysiology and powers of dispersal.

4.2.2 Interactions Between Ice and Invertebrates

Information about the interaction between river ice and invertebrates is limited. Surface ice probably has little direct effect except where it fuses with the substrate. During break-up, Logan (1963) found few organisms frozen in the ice, and floating surface ice did not appear to increase invertebrate drift. The attenuation of light by ice and snow cover (see Section 2.8.7), and low temperatures, limit attached algal production and food for invertebrates described as "scrapers" by Cummins (1974). Clifford (1978) reported minimum flows and temperatures but no oxygen depletion (see Section 3.9 for general discussion of dissolved oxygen levels) under a five-month ice cover in an Alberta brown-water stream. Most insect orders were represented, chironomids were dominant and fish were scarce. During winter, drift densities were low, most taxa exhibited little growth, there was no reproduction, and no new generations emerged. Secondary production under ice, although not specifically studied, could be negligible or negative depending on the relative magnitude of the

growth and mortality rates. The St. Lawrence River in winter as examined by Mills *et al.* (1981) represents a rather special case because of its size, the influence of Lake Ontario, and anthropogenic influences. Algal biomass was low in winter, zooplankton biomass about one-tenth algal, and beds of detrital material derived from macrophyte dieback supported the richest invertebrate communities. Molluscs dominated the benthos near Lake Ontario but were replaced by coarse particle feeders downstream where current velocity increased. Again, although not addressed, secondary production can be minimal in winter and would be supported mainly by biomass accumulated the previous summer.

Subsurface ice poses the threat to benthic invertebrates of being frozen within the ice and dislodged when anchor ice shows cyclic formation and release. Brown *et al.* (1953) found no evidence that anchor ice changed the density or composition of the benthos in the West Gallatin River, Montana (Table 4.1). The ice was described as spongy, and the nymphs of mayflies, stoneflies, and the larvae of caddisflies and midges emerged from enclosure in ice apparently unharmed. Logan (1963) found a few dead limnephilids where surface ice had frozen to the bottom at the stream margin. Benson (1955) found an average of ten invertebrates in detached square-foot samples of anchor ice. All were alive. The invertebrate fauna of drifting ice was similar. Hynes (1970) suggested that the effects of river ice are less severe than expected because much of the fauna has only to move down a very short distance into the substrate to avoid the risks. Bradt and Wieland (1981) compared the macroinvertebrate fauna of a Pennsylvania creek over a mild and a severe winter. They concluded ice scour at break-up may have reduced the densities of *Hydropsyche* and *Ephemerella*. Their results might have been influenced by different patterns of rainfall and, presumably, flows between the two winters.

Surface ice that freezes to the substrate, allowing heat loss and freezing of the stream bed, may be a more serious problem. Brown *et al.* (1953) found all the organisms dead in thawed samples from sites where surface ice had frozen to the bottom and no free water was associated with the ice or frozen substrate. Aiken (1968) concluded that crayfish (*Orconectes virilis*) could not stand freezing and any caught in the shallows where sediments froze below the surface would die. However, because mature crayfish migrate to deeper water in the fall, the population survives. Andrews and Rigler (1985) reviewed the literature on survival of benthic invertebrates in frozen sediments and concluded that many species could survive freezing at temperatures as

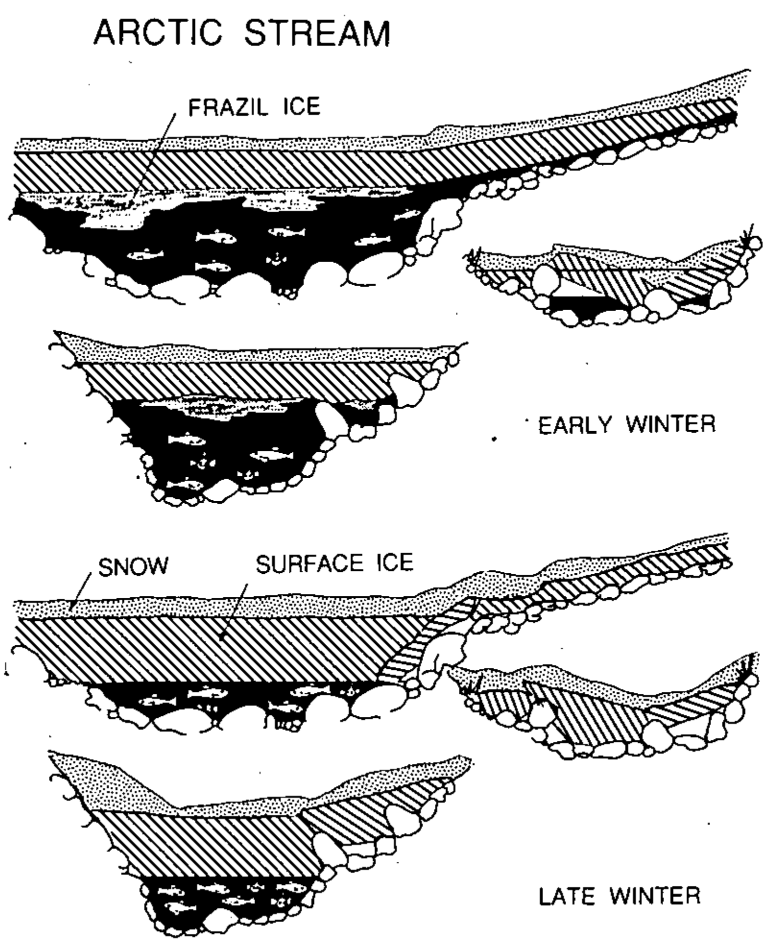
Table 4.1 Summary of biological effects of river ice.

Ice Type	Description	Detrimental Effect	Beneficial Effect	References Detrimental	Beneficial
Sheet, surface	Seals stream from outside ice	Deoxygenation	Warm area under ice for mating	Mosevich (1948)	Flannagan and Cobb (1983) Clifford (1969)
		Reduces light penetration	Provides cover, protection from predators	Shadin (1956) (Hynes)	
			Insulates water from low air temperatures		Hynes (1970)
Sheet, surface frozen into substrate	Solid wall of ice	Kills animals by freezing	Resistant life stages survive	Clifford (1969) Brown et al.(1953)	Clifford (1982)
		Moves substrate and animals on thawing	Mobile animals may move to unfrozen section	O'Donnell (1954)	
		Deposits sediments			
		Removes algal coating	Removes algal coating	Maciolek and Needham (1952)	Hynes (1970)
Anchor	Ice forms substrate - not solid often forming and melting in diurnal cycles	Moves substrate	Generally does not kill animals	O'Donnell et al. (1954)	Brown et al. (1953)
		floats material off bottom	provides cover	Hynes (1970)	
		Diurnal type causes wide change in flow	More drifting food for fish	Maciolek and Needham (1952)	Waters (1972)
		Increases "drift"		Waters (1972)	
Frazil	"Slush" in suspension	May scour bottom (see above)	May clean bottom (see above)		

low as -8 and -18°C. They believed some of the mortality they observed in Char Lake invertebrates could be attributed to sample processing. For the harpacticoid *Attheyella nordenskioldii*, they suggested winter freezing had the advantage of synchronizing the life-cycle and even increasing production. Olsson (1981) examined the overwintering of benthos in the ice and frozen sediments of Vindelalven in northern Sweden. Living Nematoda, Gastropoda, Sphaeriidae, Oligochaeta, Hirudinea, Isopoda, Trichoptera and Chironomidae were found in the ice and frozen sediments. *Asellus aquaticus* did not withstand freezing. Olsson suggested that the composition of the substrate may influence survival of animals enclosed in ice. One advantage to being frozen in the substrate is an absence of predation. In a later study (Olsson, 1988), this was investigated for the snail *Gyraulis acronicus*. Snails frozen in the shallows had a higher survival rate and greater biomass and fecundity than snails in the unfrozen sublittoral. Because initial mortality is high, the freeze-in strategy is only advantageous if the snails remain frozen long enough to benefit from the subsequent period of reduced mortality.

The physiological literature on low temperature tolerance of invertebrates deals mostly with terrestrial species, but Oswood *et al.* (1991) and Irons *et al.* (1993) reviewed the topic of overwintering of freshwater benthic macroinvertebrates. For species that cannot escape overwintering in frozen sediments, there are two options: avoid freezing by supercooling, or tolerate freezing and thawing of tissues. Invertebrates in frozen substrate or surface ice show a range of tolerance. Chironomidae generally exhibit high rates of survival, which may explain their increasing dominance northwards. Empididae also have a high tolerance to low temperatures and can survive being frozen in the substrate (Irons *et al.*, 1993). Ephemeroptera, Plecoptera and Trichoptera are uncommon in frozen substrates and may move to avoid being frozen (Olsson, 1983). A risk appears to be associated with supercooling in aquatic organisms, namely ice nucleation from surrounding ice. Organisms are, however, protected from extreme temperature in the substrate. In this context, Oswood *et al.* (1991) report reduced supercooling capabilities for Alaskan stream invertebrates. They also note that laboratory results often do not reflect field observations; survival in the field is generally higher because of the different conditions of freezing and thawing.

The effects of ice on winter discharge (see Sections 2.3 and 2.4 for discussion of hyraulic and hydrologic effects of river ice) are likely to be as important, if not more so, than freezing. Craig (1989) makes the point that even small increases in the thickness of surface ice and concomitant decreases in the amount of free water can have a pronounced effect on available habitat (Figure 4.1). Maciolek and Needham (1952) stressed the variability of winter flows caused by the daily cycle of anchor ice formation and consequent floods, diversions, and scouring. A more detailed description of the growth of ice and its



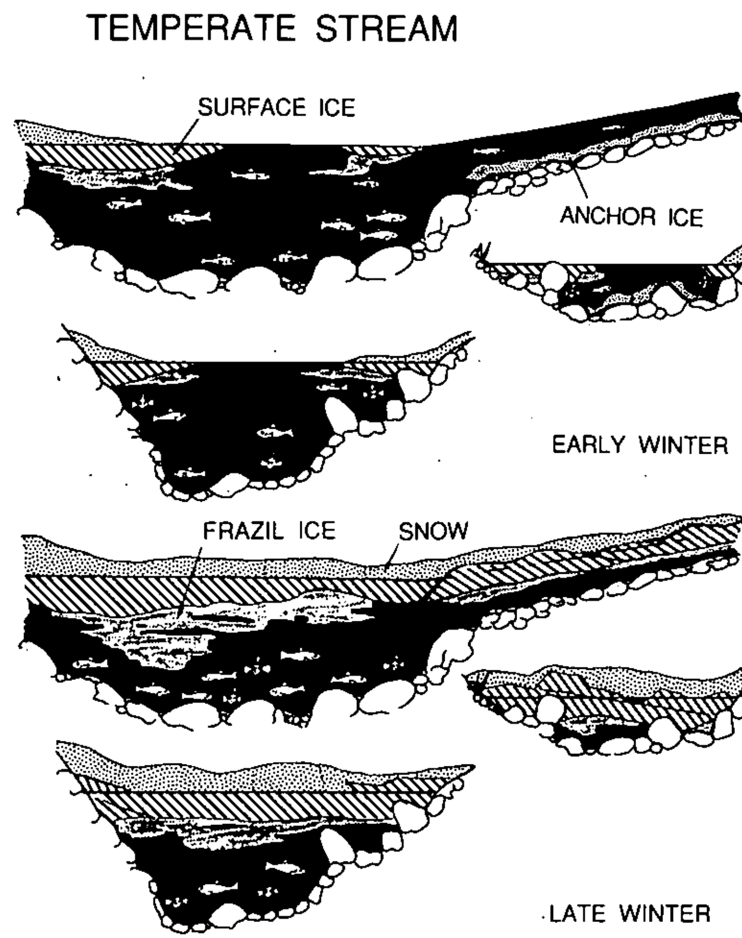


Figure 4.1 The change in available habitat in an arctic and a temperate stream riffle and pool as winter progresses. In the arctic stream the riffle may contain no free water in late winter and fish are confined to small refugia in deep pools or refugia maintained by groundwater discharge. The temperate stream maintains more habitat but open sections in riffles may generate large amounts of frazil ice which restrict the habitat available to fish and may contribute to ice dam formation with subsequent effects on discharge.

potential effects on stream flow and conditions in the hyporheic zone is provided by Blachut (1988). The hyporheic zone provides a winter refugium for many stream invertebrates (Williams, 1981; Irons *et al.*, 1989, 1993), protecting them from both freezing and flood scour. It is also the habitat in which salmonid eggs and alevins are found (see below). The critical condition for heat loss and freezing of interstitial water in the substrate is continuous ice contact between the surface and the stream bed (Blachut, 1988). Once contact is established the substrate has lost its protective layer of water and heat loss accelerates.

In contrast to the invertebrates so far discussed, the winter stoneflies (*Plecoptera*), at least when occuring in streams with marked temperature differences between summer and winter, aestivate during summer and grow during fall, winter and spring. Stewart and Stark (1988) reviewed the life history of winter stoneflies and suggest that summer diapause may be a mechanism to avoid high summer water temperatures, to synchronize nymphal growth and adult emergence, to synchronize growth of nymphs with accumulation of leaves in streams, and/or to avoid predators. Whatever the case, these animals, which can occur in very high numbers (Power, personal observation, and see Donald and Mutch, 1980), may provide considerable secondary production, under the ice in many Canadian streams. These animals are apparently well adapted to ice. Clifford (1969) observed adults, and Flannagan and Cobb (1983) observed copulating adults on the underside of river ice (Table 4.1). Flannagan and Cobb (1983) suggested that this space between water and ice cover, almost 5°C warmer than the outside air, provided a microhabitat to protect these insects from harsh weather conditions and predators (see also Section 2.2.7). In March, 1992, Cunjak observed numerous winter stonefly nymphs (last instar *Allocapnia sp*) moving about easily in the interstitial spaces of frazil ice forming a 'hanging dam' in a pool in the Northwest Miramichi River (Figure 4.2). These may have been preparing to emerge. In contrast, the eggs of middle-alpine stoneflies such as *Arcynopteryx compacta* have adapted to severe winter climates and freezing by entering diapause during winter (Lillehammer, 1987).

4.2.3 Conclusion

In summary, there is insufficient information to draw meaningful conclusions about the effects of ice on invertebrates. What little evidence there is suggests that stream benthos production is not seriously impacted by ice under natural environmental conditions. Effects observed seem to be part of the normal seasonal cycle of events to which species are well adapted. Temperature, rather than ice, limits invertebrate production in winter.

4.3 FISHES

4.3.1 Background

The study of fishes in running waters during winter has been given much attention in the past two decades (e.g., Hartman, 1965; Power and Coleman, 1967; Chapman and Bjorn, 1969; Hunt, 1969; Bustard and Narver, 1975; Belzile *et al.*, 1982; Butler, 1982; Rimmer *et al.*, 1983, 1984, Cunjak and Power, 1986 a, b; Chisholm *et al.*, 1987; Swales *et al.*, 1986, 1988; Cunjak, 1988; Veselov and Shustov, 1991; Cunjak and Randall, 1993). However, our understanding of the impacts of river ice on freshwater fishes remains poor despite the acknowledged need for its investigation over 50 years ago (Hubbs and Trautman, 1935). Much of what is known is anecdotal and is based on isolated or regional phenomena; hence, it is restrictive in its applicability over a wide range of lotic environments.

The following review attempts to summarize the state of our understanding on the interrelationships of river ice and freshwater fishes, both detrimental and beneficial. Although as many species as possible will be discussed, the emphasis will be on stream salmonids. This is a consequence of the research bias directed at these economically important species and because of the prevalence of members of this family at latitudes and altitudes where the most severe winter conditions are often realized.

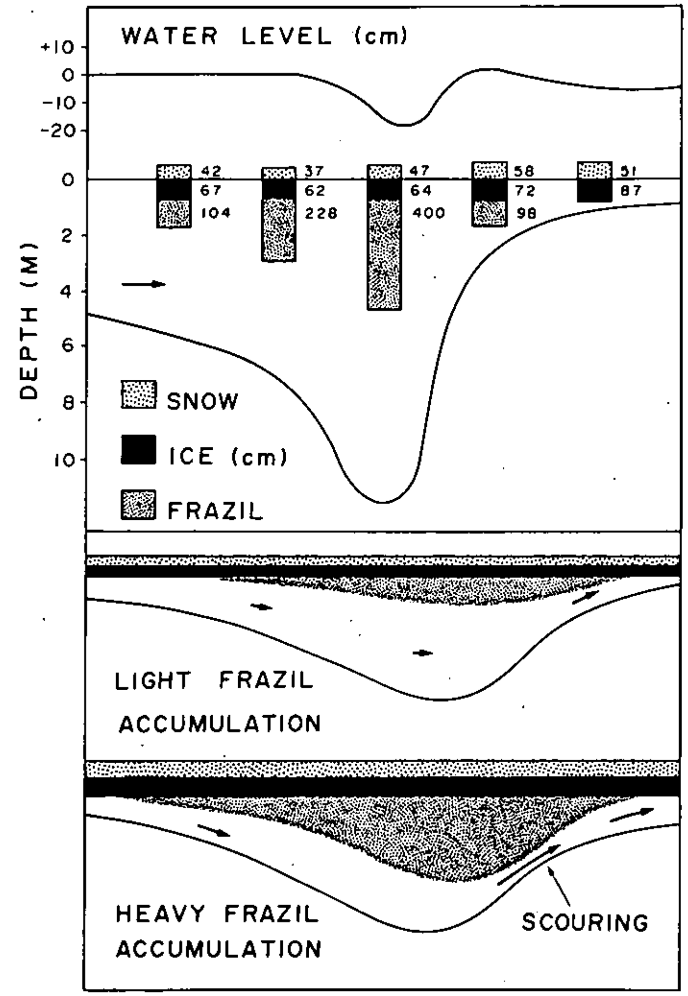


Figure 4.2 Upper Panel: Schematic of conditions in Big Hole Pool, Northwest Miramichi River, March 4, 1992, showing snow, ice and frazil accumulations at 50 m intervals along the pool. Water levels in the core holes were measured against the ice surface.

Lower Panel: Schematic of the effect of different accumulations of frazil ice on pool habitat. The amount of frazil depends on 1) the ice free area upstream, 2) heat balance (see Section 2.8.2), and 3) turbulence and discharge.

4.3.2 Effects of Subsurface Ice

Subsurface forms of ice (i.e., anchor ice and frazil ice; see Sections 2.2.2 and 2.2.3) are generally considered to be the most detrimental to fishes. Tack (1938) reported that ice crystals (frazil?) plugged the mouths and gills of trout, which led to suffocation and, eventually, to death. Such occurrences are probably rare, however. Substrate scouring, dewatering of stream sections, and freezing of redds are more common causes of mortality that are largely attributable to subsurface ice. The stream conditions often preferred for spawning by salmonid fishes (i.e. swift-flowing, shallow, gravel and cobble reaches) are also conducive for anchor ice formation. If anchor ice persists, then the growth of solid ice into the redds may be accelerated (Walsh and Calkins, 1986) and direct mortalities may result. Benson (1955) believed that brook trout (*Salvelinus fontinalis*) eggs in a Michigan river were buried below the influence of anchor ice. He speculated that developing trout embryos were most vulnerable, rather, at the time of emergence if anchor ice precluded the upward movement of sac fry. However, Reiser and Wesche (1979) recorded the freezing of eggs of brown trout (*Salmo trutta*) buried 15 cm in the substrate of a Wyoming stream even though they were covered by 12-20 cm of flowing water.

The cyclic nature of subsurface ice formation and dispersion in streams may indirectly impact upon stream fishes (Table 4.1). In Convict Creek, California, Maciolek and Needham (1952) found that brown trout and rainbow trout (*Oncorhynchus mykiss*) had consumed eyed-trout eggs dislodged from redds by the scouring action of anchor ice during its release and downstream movement. Bams (1987), in reporting on ice conditions in the Nechako River, British Columbia, identified the potential negative influences of subsurface ice on incubation-related processes. He noted that frazil and anchor ice accumulations could block the entry of oxygen-bearing water into the gravel beds where developing embryos were located (see also substrate discussion in Section 2.7.3). Maciolek and Needham (1952) and Needham and Jones (1959) reported on the mortalities of trout, by suffocation, after the dewatering of stream side-channels and braids. Trout had moved into these sections as river stage increased behind subsurface ice accumulations and then become stranded when water levels subsequently receded following detachment of the ice from the substrate.

The scouring of the stream bottom and associated high-velocity flows accompanying detachment of subsurface ice-jams (see Sections 2.5.7 and 2.7) may indirectly affect juvenile and adult stream fishes by displacing them from resting positions. Some species, such as blacknose dace (*Rhinichthys atratulus*), overwinter in a torpid or semi-torpid state beneath coarse substrates (Cunjak and Power, 1986b) and would be susceptible to dislodgement and injury if the high-velocity flows (during the release of ice-dams) were of a force sufficient to move large substrate particles (Section 2.4.7). Even fishes such as cottids, Atlantic salmon (*Salmo salar*) and steelhead trout, which are relatively active but which overwinter beneath rocks in riffles and runs (Hartman, 1965; Rimmer *et al.*, 1983; Cunjak, 1988), can be dislodged during high velocities in winter because station-holding capabilities are greatly reduced at low water temperatures (Rimmer *et al.*, 1985). Elevated river discharge during winter can displace frazil ice and developing eggs of tomcod (*Microgadus tomcod*) significant distances downstream (Fortin *et al.*, 1992). Such displacements of fish and eggs can have important effects on the productive capacities of streams, particularly if mortalities are the result.

4.3.3 Effects of Surface Ice

Surface ice, often in conjunction with anchor and/or frazil ice formations, can have deleterious impacts on fish production in streams. Dewatering of entire stream channels behind dams of accumulated subsurface and surface ice can result in fish mortalities. Where solid-surface ice cover persists for long periods, especially with a snow cover to reduce light penetration (Section 2.8.7) and primary production, reduced levels of dissolved oxygen can result (Section 3.9), limiting local production of fishes to those species tolerant of low oxygen levels.

Probably the most serious impact on stream fishes from surface ice occurs during ice break-up. Ice-dams, increased river discharge, and scouring of the stream bottom combine to make this period severely taxing to aquatic organisms. As noted in Section 2.5.7 flow velocities in excess of 5 m/s are not uncommon during break-up surges (e.g., Calkins, 1989). These strong current velocities could conceivably dislodge shelters and injure stream fishes such as dace, cottids and salmonids which overwinter in the substrate. Calkins *et al.* (1989) were able to document the movement of boulders as large as 70 kg in a Vermont stream during a spring ice-run. Flooding and ice-dams during this period can result in mass strandings of riverine species that

eventually suffocate as the water levels recede. For example, a late winter thaw in the Grand River, Ontario, in 1984 resulted in ice break-up and flooding so that thousands of fish (mainly carp, *Cyprinus carpio*) were stranded on the lands of a local golf club (Kitchener-Waterloo Record, Feb. 21, 1984).

4.3.4 Oxygen Depletion Beneath Ice

Anoxic conditions in rivers beneath ice cover, although not common phenomena, have been documented from such geographically distant locations as the River Ob in Siberia (Hynes, 1970) and rivers in Canada's north (Whitfield and McNaughton, 1986; Section 3.9). Some fishes may be able to emigrate from seasonally intolerable oxygen conditions (e.g., Magnuson *et al.*, 1985), whereas developing eggs and embryos cannot and may perish if the anoxic waters enter the gravel. Bams (1987) noted that reduced intergravel water flows (which carry dissolved oxygen to developing eggs) could result from surface ice formation that increased water depth while reducing stream velocity. In the Morice River of interior British Columbia, Bustard (1986) found a winter kill of >100 juvenile salmonids in two ice-covered, side channel pools. He attributed the mortality to low dissolved oxygen levels (<1 ppm) where aeration was precluded by low water levels and surface ice.

Problems of oxygen depletion may be widespread in northern rivers. Schreier *et al.* (1980) reported late winter depressions in dissolved oxygen (D.O.) at all stations in two Yukon rivers. They attributed this to an increasing groundwater component in the residual flow as winter progressed, and to biotic respiration. The lowest values reported (approximately 4 mg/L) may not have been lethal, except under extreme conditions, as in the Sagavanirktok River (Adams and Cannon, 1987; Schmidt *et al.*, 1989). Bendock (1981) reported D.O. values ranging from 0.6 to 5.6 mg/L in overwintering sites in the Colville River.

4.3.5 Restrictions Imposed by Variable Stream Environments

In arctic (and some continental, interior) streams and rivers, where winter conditions are extreme, the problems associated with ice formation are unique, although even less is known relative to the situation in temperate latitudes (Harper, 1981). Craig and Poulin (1975) reported on the limited distribution of arctic char (*Salvelinus alpinus*) and arctic grayling (*Thymallus arcticus*) in Alaskan rivers during winter when entire streams became frozen solid except for small overwintering areas where perennial springs precluded ice formation. Craig (1984) documented similar situations for other Beaufort Sea drainages where char, ciscoes and whitefishes occurred. He noted that those rivers lacking springs (i.e., overwintering areas) did not support anadromous fish populations. In contrast, Armstrong (1986) noted that some spring-fed tributaries that maintained flow all winter were abandoned by grayling because these streams formed extensive frazil ice, which grayling apparently could not tolerate. West *et al.* (1992) used radio telemetry to track the annual movements of arctic grayling in three North Slope Alaska rivers. They observed a well established migratory circuit between overwintering sites which offered deep pools associated with springs and distant, shallow feeding areas that could only be occupied for brief periods in summer. Other species such as juvenile coho salmon (*Oncorhynchus kisutch*) and sculpins (*Cottus cognatus*) avoided the frazil ice by inhabiting the actual springs. Hence, northern streams are limited in their productive capacity primarily because of ice conditions that prevent the use of all available habitats. These restrictions in available river habitat, due to extreme ice conditions, can have other consequences. Schmidt *et al.* (1989) found that ice conditions in an Alaskan river (Sagavanirktok) resulted in crowding of ciscoes and whitefish (*Coregonus spp*). This was believed to be responsible for the near-anoxic conditions and subsequent mortality in pools where the fishes were overwintering. Johnson *et al.* (1982) found that surface ice excluded up to 60% of the cross-sectional area of a brown trout stream in Wyoming. They speculated that preferred wintering areas would, therefore, be reduced, forcing trout to move to more favourable locations. Cunjak and Randall (1993) similarly suggested that winter movements and site fidelity of juvenile Atlantic salmon were related to habitat suitability, often a function of localized ice conditions.

In summary, both surface and subsurface ice can seriously affect the habitat of fishes in rivers. Mortality, emigration or displacement of fishes of all life-stages are often the result of severe ice conditions, through the action of damming, scouring, associated flooding or direct freezing (Figure 4.1). In arctic latitudes, where the climate is most severe, the effects of river ice and its restriction to fish production are extreme.

4.4 OTHER VERTEBRATES

Other vertebrates, aquatic as well as those in some way dependent on aquatic systems (e.g., for food or shelter), are also affected by river ice. Frogs such as the northern leopard frog (*Rana pipiens*) and the green frog (*R. climitans*) are known to hibernate in the bottoms of streams and rivers, usually in a torpid condition (Bradford, 1983; Cunjak, 1986). As was described for fishes that overwinter beneath stream substrates, these frogs may be susceptible to the action of ice (subsurface and surface) either by freezing, or dislodgement and subsequent injury or predation.

Ice cover precludes feeding by avian predators of aquatic species (e.g., kingfishes, herons and mergansers) and it is likely that their migrations may be timed to such events. Hazzard (1941) suggested that the extent and duration of surface ice on Michigan rivers directly influenced the amount of predation on stream-dwelling trout by fish-eating ducks. As lower sections of rivers begin to freeze over, birds such as American mergansers concentrate in large numbers in open, headwater streams and spring-fed tributaries where a flock of 100 mergansers has been estimated to consume 7000 trout in a fortnight (Salyer and Lagler, 1940). Early freeze-ups may impact deleteriously on such animals if the necessary energy stores for migration have not been accumulated and if open water is no longer available for continued feeding.

Mammals such as otters and minks, also dependent on aquatic organisms for food, can be similarly affected, although ice-gaps that often form beneath the surface ice in streams can accommodate movements and provide breathing space for these animals (Calkins *et al.*, 1989; see Section 2.2.7). The inundation of flood plains because of ice-dams or the high discharge associated with ice break-up can have serious consequences to terrestrial organisms living in these parts of a river drainage (e.g., small rodents).

4.5 BENEFICIAL ASPECTS OF RIVER ICE

Not all impacts of river ice are detrimental to stream organisms and, hence, production (Table 4.1). Surface ice and snow accumulation can form an insulating layer, especially in conjunction with sub-ice "air gaps" (e.g., Calkins and Brockett, 1988; Section 2.2.7). This prevents excessive radiant heat loss from rivers and maintains water temperatures slightly above freezing. This is especially true in lentic stream sections (e.g., pools, steadies and riverine ponds), where surface ice formation is most common. In drainage systems such as those found in Newfoundland, where ponds and lakes are distributed throughout the basin, the result is a series of "warming stations" restricting excessively low water temperatures. These conditions restrict the formation of subsurface ice, with its associated negative impacts on fish production (see above), and minimize substrate freezing (Calkins *et al.*, 1989). The sub-ice "air gap" has other benefits besides insulation. This micro-habitat is necessary for the successful emergence of winter developing stoneflies. Semi-aquatic mammals (e.g., otters, muskrat) use these spaces for moving long distances beneath ice sheets. The occurrence of such air-gaps is largely dependent on the availability and distribution of boulders and large woody debris to support the ice sheets above the water surface (Calkins *et al.*, 1989).

Shore-bound ice and surface ice sheets are often used as overhead cover by stream fishes, especially salmonids (Needham and Jones, 1959; Maciolek and Needham, 1952; Cunjak, personal observations) (Figure 4.3 illustrates that during the early stages of freeze-up, fish may continue to maintain stations in riffles in sheltered locations in mid channel or use the surface shelf ice as cover. With the formation of anchor ice, water velocities increase as the channel is constricted and fish either leave the riffle or shelter in low velocity areas under the shelf ice.) Surface ice sheets along river banks typically show reduced water velocities beneath as most of the flowing water is diverted toward midstream "leads". Stations beneath ice cover would be favoured by stream fishes because of the lower energetic demands to maintain positions, an important criterion for winter habitat choice among stream fishes (Cunjak and Power, 1986a). Such conditions would enhance survival and residence in streams and, ultimately, production.

The climate associated with regions where ice forms on rivers provides an indirect benefit to stream productivity. Deciduous leaves which were frozen and dried support more species of aquatic hyphomycetes and a higher production of fruiting bodies (Bärlocher, 1992). Riparian leaf-fall may experience several freezing episodes before entering streams as allochthonous matter. Such leaves will be preferentially colonized by fungi and enter the invertebrate food chain more quickly.

The life-cycle of the Atlantic tomcod, *Microgadus tomcod*, is closely linked with frazil ice-formation in rivers such as the St. Lawrence River (Anon 1985) and its tributaries (Laramée and Fortin, 1982). Spawning by this species occurs in the vicinity of "hanging" frazil ice dams, beneath ice sheets, and downstream of rapids in January and February. Subsequent development of the eggs takes place within the interstices of the frazil ice mass where it is believed the embryos are relatively protected because of the physical qualities of this ice form (Y. Mailhot, M.L.C.P., Trois Rivières, P.Q., personal communication). More recently, Fortin *et al.* (1992) have shown that the spatiotemporal distribution of tomcod eggs was related to interannual variations in river discharge and air temperature, specifically, their influence on the position of frazil ice and associated eggs.

Ice can also influence predator-prey dynamics in running waters. Avian and mammalian predators of fishes such as kingfishes and herons are unable, or at least restricted in their ability, to feed in streams and rivers after formation of an ice cover beneath which fish can hide. Winter survival of fish is thereby enhanced, especially important for poikilotherms as low water temperature (which directly affects metabolic rates in fish) may make predator avoidance more difficult in winter. Finally, the scouring action of anchor ice during its daytime release has been found to dislodge bottom organisms and eggs of trout, which are subsequently preyed upon by stream-dwelling brook and rainbow trout (Maciolek and Needham, 1952).

4.6 ANTHROPOGENIC CONSEQUENCES AND COMPLICATIONS

The preceding review of the impacts of river ice on secondary production was based on natural situations in the environment. However, human activities within or adjacent to river systems can, and often do, result in different impacts and unique complications with respect to ice conditions. Some of these will be discussed in the following paragraphs together with their implications to production and the management of lotic systems.

Hydroelectric developments alter inherently the natural hydrologic regimes of rivers, though the impacts, as they relate to winter conditions, are seldom investigated. In his report on the proposed ALCAN hydroelectric project on the Nechako River in British Columbia, Bams (1987) identified a number of potentially deleterious

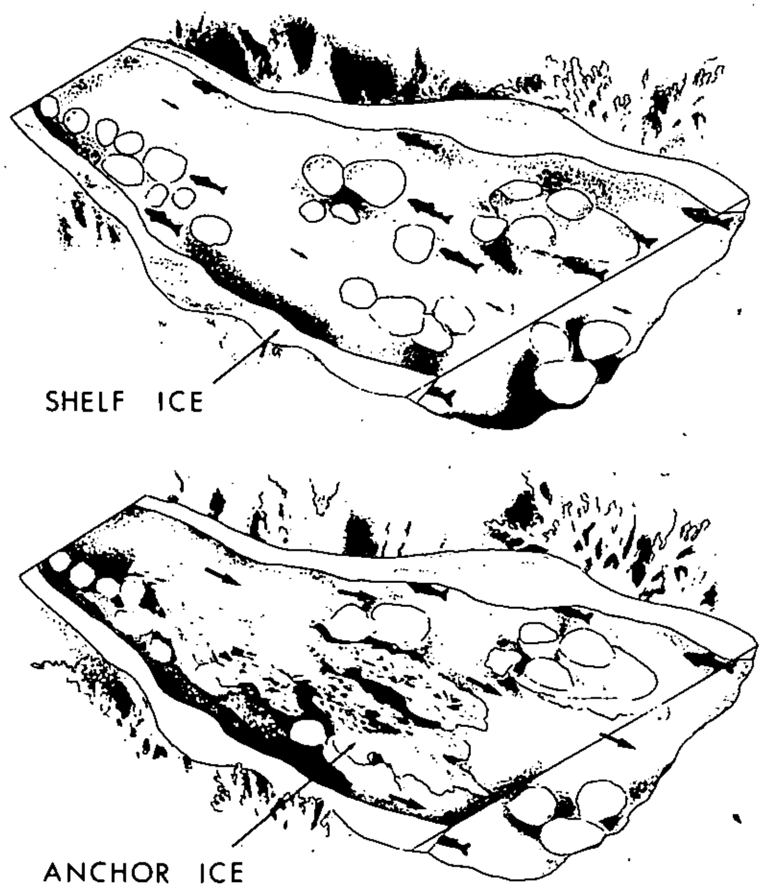


Figure 4.3 Effects of freeze-up on fish

impacts on chinook salmon populations that he related to ice conditions. He suggested that anticipated reductions in accumulated thermal units in winter as well as reduced intra-gravel flow would enhance ice formation and result in freezing of redds. This could affect incubation time, development, and survival of eggs and alevins in redds. McNeil (1966) showed that high mortalities among pink and chum salmon embryos and alevins in Alaskan streams resulted from freezing, with most mortalities occurring when river discharge was lowest. Curry *et al.* (1992) cautioned that river level reductions for hydroelectric development of the Nipigon River in Ontario could result in altered groundwater flow to brook trout (*Salvelinus fontinalis*) incubation habitat and in probable mortalities of eggs and alevins due to freezing. Since the extent of ice cover often increases with reduced flow (typical of hydroelectric diversion schemes), there are real risks of freezing of redds and other negative impacts associated with increased ice cover (see preceding discussion). Hawthorne and Butler (unpublished) also identified the deleterious effects of water releases during winter by hydroelectric operations that enhanced ice formation below discharge structures. Such unnatural releases of water can dislodge torpid or hibernating aquatic organisms (e.g., fishes) which can then be preyed upon or injured when swept downstream.

Winter demands on water resources such as those posed by alpine ski operations in New England are ever increasing (D. Calkins, personal communication). Subsequent lower water levels and reduced discharge could exclude aquatic species from available overwintering habitats because of increased surface and subsurface ice formation. For example, West *et al.* (1992) emphasized the susceptibility of North Slope (Alaska) grayling to habitat damage by water abstraction and drainage changes because of the shallowness of rivers and the extremely limited overwintering habitat. Chisholm *et al.* (1987) suggested that water withdrawal schemes will vary in their impacts to lotic systems in winter depending on the elevation of stream reaches, mainly because of the differential build-up of ice at different altitudes.

Lasko (1987) noted problems of oxygen depletion (<4 mg/L) beneath ice cover in Finnish rivers, especially where pollution loads increased biotic oxygen demands. To correct the problem, aeration dams were successfully used.

Because many riverine species are poikilotherms, water temperature directly controls metabolic rates and growth. Factors that alter the thermal regimes of streams will, therefore, alter rates of production.

Man-made structures such as dams, causeways, and bridges across rivers often constrict waterways and hold back ice above the obstructions longer than under natural circumstances. Protracted periods of ice cover would reduce the available thermal units available to aquatic organisms during winter and spring. Since animals and their life-stages (e.g., spawning, development) are finely tuned to precise and dependable environmental cues by natural selection, such abrupt changes can upset the critical balances of nature. Obviously, more research in winter is required to address such problems in order to properly manage aquatic environments.

4.7 DISCUSSION

A broad overview of the effects of river ice on biota must consider latitude, altitude, local geology, climate, and exposure, as well as post-glacial history and faunal evolution. Ice impacts are likely to be most obvious as latitude, altitude, and distance from the ocean increase (but see Chisholm *et al.,* 1987).

In temperate and coastal-maritime regions, where a proportion of the winter precipitation falls as rain and there are periodic melt episodes, low-flow conditions are less likely, though ice scouring, in association with floods, is capable of reducing standing stocks of fish and invertebrates (e.g., Elwood and Waters, 1969; Erman *et al.,*, 1988). Northwards, at higher elevations, and inland, more or all of the winter precipitation is locked up in snow, and the duration of ice cover increases. In regions of discontinuous permafrost, near-surface run-off is impaired once the active layer freezes, and stream discharge is variously reduced, depending on lake storage, deeper groundwater sources and the severity of winter. The most severe flow restrictions occur in arctic regions with continuous permafrost. Apart from exceptional groundwater discharges, flows reach very low levels or cease altogether in low order streams. Evolution and selection must be considered in answering the question of whether river ice affects secondary production. Temperate and arctic faunas have evolved to deal with a suite of seasonally predictable conditions and one would expect that it is only when these conditions exceed the normal range that mortalities occur.

Attempting to integrate our ideas about possible impacts of river ice on secondary production with theoretical ideas about how lotic systems operate may provide some insight. Ecological of

consideration include:

1) The river continuum concept (Vannote *et al.*, 1980; Naiman *et al.*, 1987);
2) The abiotic-biotic continuum concept (Zaleski and Naiman, 1986; Schlosser and Ebel, 1989);
3) r - K strategy and life-history theory (MacArthur and Wilson, 1967; Stearns, 1977); and
4) Intermediate disturbance hypotheses (Connell, 1978; Fisher, 1983; Miller and Stout, 1989).

These frameworks are not mutually exclusive nor geographically isolated. All have been used to explain the community composition and dynamics of running water ecosystems either in general or in specific instances. Added to the above are ideas and techniques worked out in relation to the in-stream flow incremental methodology and the estimate of weighted usable area (Bovee, 1982; Bovee and Cochnauer, 1977). In streams and rivers, the rate at which habitat is lost increases rapidly as flows decline below about 40% of the mean annual discharge (Tennant, 1976). As useful and interesting as the aforementioned techniques and concepts are, they are probably not developed sufficiently well to help answer the question(s) discussed in this review. Clearly more detail is needed before the role of ice in lotic ecosystems can be fully appreciated.

5

SUMMARY

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The objective of the preceding chapters was to provide the reader with a basic background to river-ice types and covers, and to review related physical, chemical, and biological environmental effects. Although the references cited do not represent an exhaustive cataloguing of all available information, the selected topics within each section hopefully encompass the major emerging issues related to the environmental effects of river ice. Notably, however, the list of important issues is rapidly expanding as more attention is placed on the river-ice period. This has resulted for two reasons. The first stems from a developing awareness of the broad environmental impacts that have resulted, or could result, from changes/modifications to river-ice regimes. Much of this awareness has been generated by environmental-impact assessments of some of the water-development schemes that have affected major northward-flowing rivers in cold regions over the last quarter-century.

Secondly, and partly as a result of the notoriety of the above environmental impacts, many natural-science disciplines have begun to shift more of their research efforts away from the traditional studies of the open-water period to that of ice-conditions. Such a re-direction of research could be considered a valid and needed correction to an historical overemphasis on open-water conditions. This is especially true for a country such as Canada where many rivers are ice-covered for half the year and ice effects are often responsible for the major hydrologic events. As more research has been focused on the ice-covered period, an increased awareness has rapidly developed of the significance of this previously "black-boxed" period of the year.

Given the very high probability of further development pressure on northern ice-dominated flow regimes, it is critical, for achieving effective environmental protection, to obtain an improved, and broader, understanding of the environmental effects of river ice. Such an improved understanding should help us to: better focus planning and assessment studies; define research needs and priorities; modify development plans to protect valued environmental resources; and work towards the principles of sustainable development. The following identifies a limited number of broad areas of research needs. Some are specific and others were formulated by linking complementary discussions from the various sections. For the sake of simplicity, examples are identified in a chronological order from initial ice formation to break-up.

The environmental effects of river ice are present even before a full ice cover is established. Unlike lakes, where ice formation is dominated by static-ice growth, the dynamic growth processes on rivers are much more complex and can have significantly more environmental effects. The growth of frazil and anchor ice is still a major focus of hydraulic and thermodynamic scientific study, but it is also one of critical concern to riverine habitat. While frazil ice deposited beneath an ice-cover can create a unique biological environment, the formation of anchor ice on a river substrate can destroy critical habitat zones through restrictions in inter-substrate flow, enhanced freezing, and decreased oxygen supply. Given that flow abstraction from rivers increases the level and frequency of low-flow events, and thus the probability of anchor ice formation, more attention needs to be focused on the impacts of anchor ice on stream and river habitat.

Frazil ice has also been identified as a scavenger of particulate matter within a flow system. Although this has only minor significance to, for example, overall sediment transport, it becomes much more important when one considers that many contaminants released to river systems, such as organochlorines, appear to have an affinity for small-grain suspended-sediment particles. This has special implications for any pollutant sources located at or upstream from frazil-producing reaches. If the effluents are warmer than river flows (which is typically the case), the warm-water effluent from such sources is likely to preclude the development of a complete ice-cover but permit continued frazil production (at the downstream cooled portion of the reach) and

scavenging of pollutants throughout the winter. Given the high-priority of river-pollutant studies, the scavenging by frazil ice of fine-grained suspended-sediment and other contaminant-affected particles, and their downstream transport/deposition, are definitely worthy of additional research.

Once an ice cover becomes established, its most obvious impact is on water levels. The sudden, additional resistance of an ice cover serves to elevate water levels upstream but, more importantly, also leads to decreased flow and water levels downstream. Although winter low-flow has been the focus of considerable scientific study, the emphasis has been placed on the low-flow period in late-winter produced by minimal landscape runoff. As pointed out in this report, however, the ice-induced period of low flow at freeze-up can be characterized by even lower stage and discharge than those in late winter. Furthermore, this can be a protracted period, depending on the size and hydraulic-composition of the river system. As such, it will have special environmental significance to, for example, overall water supply to riverine habitat and dilution and dispersion of pollutants. More study needs to be conducted on the frequency, magnitude and duration of such events and their broader environmental effects.

A solid ice cover effectively seals a river system and eliminates direct atmospheric-water exchanges. The effects of this on oxygen levels in lakes are well documented but are only now becoming a major focus on rivers. While these effects may be important to, for example, the survival of oxygen-sensitive fish on undeveloped river basins, they have enhanced significance for aquatic life on rivers that receive additional loadings of contaminants (e.g., nutrients, carbonaceous matter, etc.) that can accelerate the consumption of under-ice oxygen.

Modelling of oxygen levels under ice is in a state of infancy. Future refinements will require the incorporation of some important ice effects, such as the attenuation of solar radiation that drives photosynthetic production, and the re-aeration that occurs through open-water polynyi. In view of the probability that the latter are also sites of concentrated biological activity, more aquatic studies should be made of these open-water zones. More generally, further studies should be made of all the unique aquatic habitats that are created by an ice cover: under-ice air pockets being one other obvious example.

In general, winter flows under an ice cover are at the annual minimum and, hence, the aquatic system is less capable of adequately assimilating the same level of pollutant/contaminant inputs as during the higher-flow open-water period. Regulation/management is thus oriented towards ensuring that specific levels of concentration are not exceeded. As explained in Chapter 2, however, there is an incomplete understanding of the mechanisms by which dissolved or suspended substances are spread and mixed under an ice cover (i.e., through vertical diffusion, transverse mixing and longitudinal dispersion). Effective regulation/management of water quality for under-ice conditions can only be achieved with enhanced scientific efforts on under-ice mixing studies.

Low winter flows also lead to sediment deposition. Lighter, smaller-grained portions of the suspended-sediment fraction will deposit on the bed as the flow and velocity decrease. Again, since such material has a strong adsorptive affinity for aquatic contaminants, under-ice low-flow conditions are likely to dictate the spatial distribution of a major portion of the accumulated winter-load of contaminants. This places an added emphasis on the need for understanding the specific hydraulics of under-ice sediment transport, including that related to the flocculation of fine-grained particles.

While under-ice low flows may determine the winter distribution of contaminated sediments, their subsequent transport within a river system is next controlled by the spring freshet and accompanying ice break-up. As noted, however, this is the period for which it is most difficult to model even the basic flow conditions. Although it is generally recognized that break-up is capable of producing abnormally-high (compared to open-water conditions) velocities and associated bed-scour, there remains some doubt about its overall geomorphic significance compared to open-water high-flow events. On certain river systems, however, the evidence is clear that ice action creates unique bank and substrate features. Such activity also produces suspended-sediment concentrations that exceed those for equivalent discharge under open-water. All of the above sediment/geomorphic processes have strong implications for aquatic life during the break-up period. Some of the rapid changes, such as in water temperature and radiation receipts associated with dynamic break-ups, are also likely to be biologically important. However, despite the strong physical changes that could be imposed by break-up on riverine habitat, relatively few biological studies have focused on this period of intense change and activity. It may well be that for some systems break-up may be the most severe

type of physical disturbance and could form the biological set-point for the remainder of the year, imposing limits on the biological productivity and/or carrying capacity of the system.

As detailed in various parts of this report, river ice break-up can also have beneficial effects for aquatic systems. Flooding is often viewed with a negative connotation because of its economic and societal impacts. For natural systems, however, flooding is a source of necessary replenishment. Moreover, in some systems the required frequency of large-magnitude flooding can only be produced by ice-jam backwater. The dominant role of ice-jam flooding over open-water flow events has only recently been recognized in some unique riverine habitats, such as the Peace-Athabasca Delta. Attempts are now under way to create artificial ice-jam flooding of this unique ecosystem that has been impacted by almost two decades of drying resulting from changes in the flow/ice regime. While the effects of reduced ice-jam flooding are relatively obvious in an ecosystem such as a river delta, comparable but less-apparent impacts also occur within the channel of a river or stream. Reduced flooding of snye habitats, accelerated channel aggradation, and enhanced succession of shore vegetation are just three examples of environmental impacts that could result from reduced break-up and ice-jam severity. Further research is required to assess overall in-channel effects.

River ice break-up is one activity that can easily be affected by changes in flow regime through, for example, regulation, diversion, or even climate change. Changes in the natural frequency and magnitude of dynamic and thermal break-up events can have far-reaching environmental implications. To date, scientific studies have only scratched at the surface of the environmental significance of river-ice break-up.

This chapter has attempted to recap some of the most important environmental effects of river ice that were reviewed in the preceding chapters. It is in no way exhaustive and probably exhibits some of the personal biases of the authors, but, hopefully, it draws closer into focus the need for inter-disciplinary research dealing with river ice. This report forms the basis for theme presentations to be presented at an August 1993 workshop of the same name. It is also intended to be a springboard for further discussion and work in this expanding field.

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