Chapter 15

River Ice

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Abstract River ice processes have an important influence on winter hydrology of cold regions. During freeze-up excessive frazil ice production can obstruct water intakes to constrain hydro-power production, and frazil accumulations can be detrimental to fish habitat. Frazil problems may persist through winter or, alternatively, mid-winter thaws may lead to premature breakup and possible ice jam flooding. The river ice breakup period may be characterized by severe ice runs associated with ice jam formation and release, with potential impacts on infrastructure, and a high risk of flooding. An overview of river ice research undertaken in the past decade is presented, including investigations into the potential impacts of climate change on rivers in the Mackenzie Basin, observations of dynamic river ice processes such as ice jam formation and release, application of satellite remote sensing techniques for river ice characterization and the development of new hydraulic and logic based models for ice jam flood forecasting.

1 Introduction

One unique aspect of cold region hydrology is the influence of winter on streamflow behavior. Most Canadian rivers experience some ice effects each year, and in many cases the runoff events associated with river ice have produced the most extreme and dangerous flood events on record. This is because breaking river ice forms ice jams that obstruct the passage of runoff and can raise water levels far higher than those experienced for the same flows under open water conditions. Therefore, despite the fact that river ice processes tend to occur on relatively small scales (in the order of tens of kilometers) they can significantly affect basin hydrologic response in term of channel routing efficiency, with the consequent influence felt over hundreds of kilometers.

River ice can also be beneficial. For example, in many areas of northern Canada, ice bridges across rivers provide access to remote communities, and many rely on these winter crossings for essential transport of supplies and people. Even more populated communities take advantage of river ice crossings for more convenient public transportation, or for industrial trans-

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port to and from mining or lumber operations. River ice covers have also been used as convenient platforms for bridge construction or bridge foundation testing. In less populated areas, river ice jams can actually be beneficial in creating water levels sufficiently high to replenish shallow lakes and wetlands, as in the case of the Peace-Athabasca Delta (Beltaos et al. 2006a).

Recent Canadian experience suggests that climate change has already begun to influence the winter regime of northern rivers. Many northern communities are experiencing warmer weather, which in turns limits the viability of some ice roads and crossings. Climate warming may also have the potential to increase the frequency and severity of ice jam related flooding in certain Canadian regions (Beltaos 2002). It is critically important to realize that river ice processes are not only affected by basin hydrology, they can affect basin hydrology. Consequently, realistic models of the impact of climate warming on basin hydrologic response in northern regions must have a deterministic component that considers the interaction of climate, hydrology and river ice hydraulics simultaneously.

This chapter provides an overview of the unique nature of river ice processes from freeze-up, through the winter and during breakup, including a discussion of the effects of streamflow regulation on the winter regime of rivers. Emphasis is placed on breakup and attendant ice jams, which are the processes that have the most serious ecological and socio-economic impacts. A synopsis of river ice research areas undertaken as part of the Mackenzie GEWEX Study (MAGS) is also presented, including investigations of: new streamflow monitoring and remote sensing techniques; novel ice jam flood forecasting methods; climate and regulation impacts on ice-jam flooding of northern rivers; and the development of numerical models of river ice processes. A summary discussion of the potential impacts of climate change on the winter regime of rivers is also presented.

2 Overview of River Ice Processes

2.1 Freeze-up

The first stage in river ice cover development on northern rivers is water cooling. The primary source of heat transfer is convective heat loss from the water surface to the colder overlying air. Solar radiation in the daytime contributes small amounts of heat to the overall energy budget; but in the Mackenzie River Basin daily heat gain from solar radiation during the freeze-up period is nearly balanced by daily heat loss due to long-wave radiation emissions, and thus the two can generally be neglected. As the cooling period progresses, water temperatures eventually reach 0°C. However, further cooling of the water to at least a few hundredths of a degree below 0°C is necessary before the first ice formation can practically occur. This is known as “supercooling”.

The onset of freeze-up begins with the development of frazil particles (small discs of ice 1 to 3 mm in diameter) that form in the supercooled water. In the slower flow near the banks (e.g., less than about 0.1 m s^{-1} velocity), ice particles develop near the surface and accumulate to form a continuous layer of skim ice on the water surface. This skim ice effectively prevents further supercooling, and subsequent ice growth is thermal in nature. The resulting ice cover is typically termed “border ice”. Because ice formed by thermal heat exchange across the ice layer usually results in crystal growth in the vertical direction, a characteristic of thermal ice is its columnar crystal structure, easily recognizable in the “candles” of ice seen as this type of ice melts.

Frazil particles also form in the faster moving portions of the flow (away from the banks). Figure 1 traces the formation of ice in the main-flow zone of a river, starting with ice-free, above-freezing water and ending with zero-degree water carrying large ice floes. Due to turbulence, water cooled at the surface is mixed through the flow and leads to an apparent spontaneous generation of frazil particles, occurring throughout the depth (once the water temperature cools below 0°C). Individual frazil particles tend to behave in a highly adhesive fashion while in supercooled water. Adhesion is generated when the particles melt briefly as they collide with other ice particles or objects, due to the small amount of heat produced by the collision, but refreeze readily in the supercooled water. This adhesive nature of the frazil particles causes them to accumulate, forming “frazil slush” (also known as “frazil flocs”). These frazil flocs eventually reach a size at which buoyant forces overcome the ability of the flow turbulence to maintain the flocs in suspension, and they float to the water surface. Here, they fuse with other flocs to form larger elements whose unsubmerged portion freezes into the familiar “pancake ice” (also known as “frazil pans”). Some of the frazil particles or pans may also collect along the border ice. This increases the border ice encroachment on the channel, and is termed “buttering”.

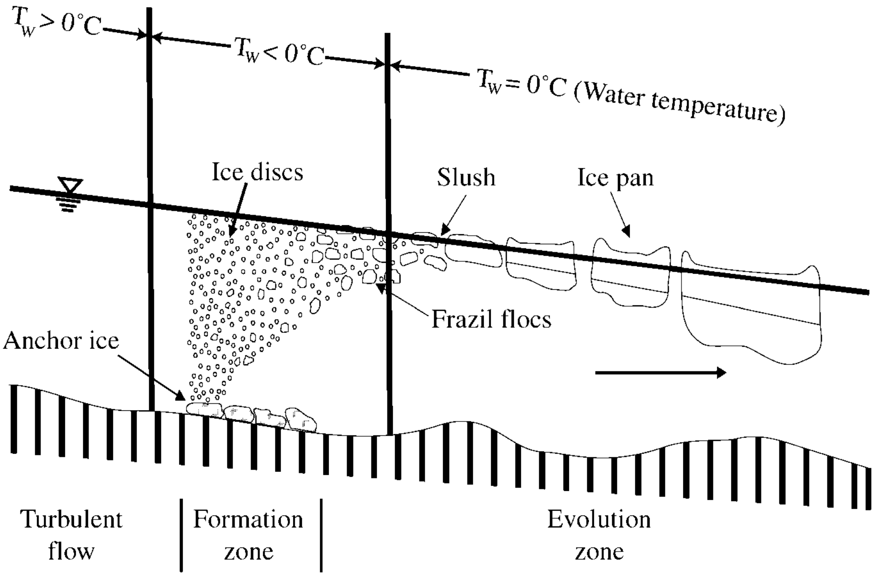

Fig. 1. Stages in river ice cover development

Turbulence also causes some frazil particles to impact the bed, and pick up small sediment particles before accumulating into large enough flocs to float to the surface. When this happens, the frazil slush layer underlying the pans may contain sediment particles. When the frazil particles adhere to very large gravel or boulders they can remain on the bed forming an ice accumulation known as "anchor ice", which releases and floats when the water is no longer supercooled.

Frazil pans float downstream on the water surface. As surface concentrations increase (both in time and in the downstream direction) the individual pans may ride up on, or freeze against other pans forming 'rafts'. When the concentration reaches about 80–90%, "bridging" often occurs. This involves a congestion of ice floes and a subsequent cessation of their movement at a site along the river. Once bridging is established, the incoming ice floes may lead to an upstream progression of the ice front by "juxtapositioning" with ice floes accumulating edge to edge on the water surface. However, if flow velocities are high enough, it is also possible that surface ice floes arriving at the ice front may be swept under the ice front and then deposited on the underside of the cover. This process is known as "hydraulic thickening". The increased thickness results in an increase in water level and a corresponding decrease in flow velocity. With a sufficient reduction in flow velocity, ice floes are no longer swept under the ice cover and the ice front can continue its upstream progression. In extreme cases, velocities may be high enough that the entire ice cover formed at the bridging site maybe swept downstream, after which bridging must again initiate before frontal progression of the ice cover can occur. Any one of these three scenarios may be observed at a given site at different times. Which of the three is to be expected at any given time is a function of both meteorological and hydraulic conditions.

As the ice front progresses upstream, either by juxtapositioning or by hydraulic thickening, the forces acting on the ice accumulation increase. These forces include the downslope component of ice weight within the ice accumulation, and the flow drag along the underside of the ice cover. These forces are resisted by the internal strength of the accumulation which, for freeze-up accumulations, is often enhanced by freezing between the individual ice floes. The forces acting on the ice cover increase as it lengthens, and when the magnitude of these forces approaches the internal strength of the ice accumulation, the ice cover is prone to collapse, or "shove", and thickens substantially as the ice front progresses upstream. The increased thickness and roughness of the ice cover after such a collapse is usually reflected in a dramatic increase in water levels. The resulting accumulation is termed a "freeze-up ice jam" or "hummocky ice cover". Normally once the accumulation has stabilized, the water between the ice floes freezes and gives strength to the accumulation, thereby inhibiting further consolidation.

2.2 Winter

Once a stable ice cover is established and cold weather persists, the solid-ice layer thickens by freezing at the water-ice interface. Where there is no slush deposit under the solid-ice layer, the original crystals grow vertically downward forming clear, columnar ice that is commonly called blue or black ice. If a slush deposit is present, the thickening process will be accelerated by the fact that a certain fraction of ice is already present and it is only the interstitial water that needs to freeze. Freezing causes expulsion of impurities, which tend to concentrate at the crystal boundaries where they can play an important role in the decay of the ice cover during the breakup period. The snow cover insulates the ice sheet and retards growth but it can also enhance growth via formation of snow ice. This is a relatively opaque layer that forms by freezing of overflow water in cases where the phreatic water surface is above the top of the ice sheet. For temperate lakes, Adams and Prowse (1981) found that the decrease in black ice growth due to insulation can be offset in the long term by the additional ice thickness created by snow-ice.

The growth of the solid-ice layer during the winter slows down as the layer becomes thicker. In many applications, the well-known Stefan formula, which is indexed by the accumulated freezing degree-days, provides a simple means for approximate prediction of the solid-ice thickness (Michel 1971). More sophisticated approaches explicitly account for various factors other than air temperature, such as solar radiation, wind speed, relative humidity, cloud cover, as well as snow depth and density (Menard et al. 2002). Values of the average maximum thicknesses of solid ice in Canadian rivers range from less than 0.3 m in the more temperate regions of Southern Canada to over 1.7 m in the Arctic (Prowse 1995). The dates on which the various maxima are attained vary significantly because of climatic differences. For example, river ice continues to grow well into April in the Mackenzie River delta (Sherstone et al. 1986), long after breakup has occurred in southern Canada.

Much higher growth rates and extreme thicknesses can be expected where thin layers of slowly moving water are continuously exposed to the atmosphere, forming aufeis, also known as icings or naleds. These are accumulations of solid ice produced by the seepage of water onto existing ice covers, and are most commonly encountered on arctic and sub-arctic rivers. During periods of runoff, icings 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 (Prowse 1995).

2.3 Breakup

The breakup of a river ice cover is triggered by mild weather and encompasses a variety of processes associated with thermal deterioration: initial fracture, movement, fragmentation, transport, jamming, wave motion and ice runs, and final ice clearance. Although several or all of these processes may occur simultaneously within a given reach, it is convenient to visualize the breakup period as a succession of distinct phases such as pre-breakup, onset, drive, and wash. During the pre-breakup phase, the ice cover becomes more susceptible to fracture and movement via thermally induced reductions in thickness and strength. At the same time, the warming weather brings about increased flows, due to snowmelt or rainfall or both. The increasing hydrodynamic forces fracture the ice cover, while the rising water levels reduce its attachment to the riverbanks. Eventually, large segments of the now fragmented cover are dislodged and set in motion by the flow. This is the onset of breakup, and is followed by the drive, that is, the transport and further breakdown of large ice sheets into smaller blocks and rubble. The onset is governed by many factors, including channel morphology, which is highly variable along the river. It is thus common to find reaches where breakup has started, alternating with reaches where the winter ice cover has not yet moved.

Invariably, this situation leads to jamming because ice blocks moving down the river in one reach encounter stationary ice cover in another reach and begin to pile up behind it, initiating a jam (Fig. 2). Ice jams can stay in place for a few minutes or for many days; they can be a few hundred meters or many kilometers long. Ice jams can cause much higher water levels than are possible under open-water conditions with the same river discharge, owing to their considerable thickness and roughness (Beltaos 1995). On many Canadian rivers, the highest water levels result from ice jams rather than from open-water floods. The wave and ice run that follow the release of a jam often dislodge and break up long sections of intact ice that they encounter; on other occasions, the stationary ice is too strong or the wave is too attenuated, and the ice run is arrested, forming a new jam. In this manner, more and more ice is broken up and carried down the river, until the final jam releases. This is the start of the wash or final clearance of ice.

Depending on hydro-meteorologic conditions, the severity of a breakup event can vary between two extremes, those of the thermal or overmature breakup and the premature breakup. The former type occurs when mild weather is accompanied by low runoff, due to slow melt and lack of rain. The ice cover deteriorates in place and eventually disintegrates under the limited forces applied by the modest current. Ice jamming is minimal, if any, and water levels remain low. Premature breakup on the other hand, is associated with rapid runoff, usually due to a combination of rapid melt and heavy rain. The hydrodynamic forces are sufficient to lift and break segments of the ice cover before significant thermal deterioration can occur. Ice jams are now the most persistent because they are held in place by sheet ice that retains its strength and thickness. This is aggravated by the prevailing high river flows, so that premature events are the most severe in terms of flooding and damages. Usually, a breakup event falls somewhere between these two extremes, and involves a combination of thermal effects and mechanical fracture of the ice. Herein, the term mechanical breakup is used to denote all non-thermal events because they are, at least in part, governed by the mechanical properties of the ice cover.

Fig. 2. Aerial and ground views of ice jams: (top) Peace River in Wood Buffalo National Park, Alberta (Photo: S. Beltaos), and (bottom) Hay River near the town of Hay River, Northwest Territories, ice floe thickness about 7 m. (Photos: F. Hicks)

In the colder continental parts of Canada such as the Prairies or the Territories, we are most familiar with a single event, the spring breakup, which is triggered by snowmelt. In more temperate regions, however, such as parts of Atlantic Canada, Quebec, Ontario and British Columbia, events called mid-winter thaws are common. Usually occurring in January and February, they consist of a few days of mild weather and typically come with significant rainfall. River flows may rise very rapidly and sufficiently to trigger breakup on many local rivers. This is the mid-winter breakup which can be much more severe than a spring event, because of the sharp rise in flow that results from the rain-snowmelt combination. The premature nature of a mid-winter breakup event almost ensures the occurrence of major jams. Moreover, dealing with the aftermath of flooding is hampered by the cold weather that resumes in a few days, while many mid-winter jams do not release but freeze in place, posing an additional threat during subsequent runoff events.

2.4 Breakup Regime of Northern Rivers

The breakup of northern rivers is triggered by spring melt and typically occurs after several days of bright sunshine that helps reduce ice thickness and strength. Mid-winter thaws and runoff events are either rare or completely unknown. Breakup is thus more predictable than on more southern rivers, and often can be anticipated days or weeks in advance, based on flow hydrographs occurring at upstream hydrometric stations. Maximum winter ice thickness is considerable, ranging from about 0.6 m on the upper Peace and Athabasca Rivers, to 1 m in the lower Peace and Slave, to over 1.5 m in the Mackenzie Delta channels.

Low water surface slope is another characteristic feature of the main rivers of the Mackenzie Basin. For instance, the slope of Peace River downstream of Carcajou drops below 0.1 m km^{-1} (Kellerhals et al. 1972). In the main channels of the Mackenzie Delta, the water surface slope is about 0.01 m km^{-1}. It is doubtful whether such flat reaches can generate sufficient driving forces to dislodge the thick ice cover under snowmelt-runoff conditions. Mechanical breakup is more likely to be generated by waves that result from the releases of upstream ice jams. Such waves can greatly amplify the driving forces (Beltaos and Burrell 2005b) and cause breakup over extensive river lengths (Gerard et al. 1984). Each wave is accompanied by an ice run that is eventually arrested by competent ice cover downstream to form a new jam. Essentially, mechanical breakup process consists of a sequence of jams and ice runs. Where this process is stalled for prolonged periods of time, thermal effects cause severe decay of the winter ice cover, resulting in a thermal event.

A key factor affecting many of the large rivers in the Mackenzie River Basin is that they are generally north flowing (e.g., Peace, Athabasca, Slave, and Mackenzie). Thus they tend to break up first in the headwaters, which are located in the most southerly portions of their respective basins. As they advance downstream, the dynamic ice runs discussed above are thus more likely to encounter a strong (undeteriorated) ice cover, since this cover is located further north. This sequence of events enhances the potential for ice jam formation as compared to rivers flowing in other directions.

2.5 Effects of Streamflow Regulation

On regulated rivers both water storage in the reservoir and flow release patterns have the potential to significantly affect river ice processes. Reservoir storage is important because it raises winter water temperatures in the downstream reach. This occurs because of the unique density characteristics of water (Ashton 1986). As with other fluids, water density varies with temperature. However, water density is maximum at 4°C, and decreases with temperatures both below and above 4°C. In deep reservoirs containing water at temperatures in excess of 4°C, the cooler, denser water is found at greater depths than the warmer, less dense water. This vertical stratification is stable because further heating of the surface layers of water only leads to reduced density in these upper layers. However, as water in a reservoir cools below 4°C, it develops an inverse temperature gradient. Initially, surface heat loss lowers the water temperature in the upper layers towards 4°C and this denser water then sinks to the lower levels. As the temperature cools the water further, the water density decreases and the colder (but less dense) water remains nearer the surface. The resulting profile is at 0°C near the surface and 4°C at the reservoir bed. Further heat loss through the winter season has the potential to cool the water through the entire depth. However, there will still be a temperature gradient until all of the water is cooled to 0°C. The temperature gradient generally persists throughout the winter, as the formation of an ice cover insulates the water from cold air temperatures. Snow accumulations on the ice cover enhance this insulating effect. Since water is typically drawn from the bottom of the reservoir, flow releases from dams are usually above 0°C throughout the winter. Note that the vertical temperature gradient does not persist once the water enters the river, due to the turbulent nature of the flow. Consequently, temperature in the river is expected to be homogeneous throughout the flow depth.

The release of warm water from a reservoir can affect a river's ice regime in three ways (see also Starosolszky 1990). First, it can inhibit the early formation of an ice cover in the upper reach of the river (near the reservoir outlet). This generally leads to a prolonged and relatively unstable freeze-up period (i.e., one prone to ice consolidation events). On rivers regulated for hydro-power production which are subject to hydro-peaking operations (highly variable discharge conditions), this unstable period may be prolonged. Second, it can limit the thermal growth of ice in the downstream channel. Therefore, thinner ice covers might be expected as compared to the pre-regulation period. Third, it could lead to the early melt of river ice in the spring and a greater tendency for thermal breakup events.

Regulation may also alter the flow hydrograph, and thence the ice regime, of the river downstream of the regulation site. For example, hydropower generation by means of large reservoirs greatly augments fall and winter flows, resulting in thick ice covers and high freeze-up water levels, which can modify the frequency and severity of spring breakup jamming (Beltaos 1997; Beltaos et al. 2006a) or promote mid-winter breakup and jams (Andres et al. 2003). Breakup flows may be either augmented or reduced, and this effect can also modify the jamming regime.

Upstream effects arise mainly from a change in the water surface profile of the river. Relatively thick ice covers form over the surface of the reservoir, and appear earlier than under natural conditions (Starosolszky 1990). The thicker reservoir ice and the low water surface slope of a reservoir promote ice jam formation during breakup, not only within the main river but also at the mouths of tributaries located within the reservoir reach.

3 Monitoring River Ice Processes

3.1 Overview

Determining the areal extent and nature of the ice cover is a key objective of many river ice monitoring programs. Particularly during breakup, monitoring the development of open water leads, major ice movements, ice runs, and ice jam formation and release is often done with the aid of aerial reconnaissance flights, documenting with digital video or still camera. Water level monitoring is also generally undertaken, using both manual measurements and automated monitoring networks (e.g., Robichaud and Hicks 2001) with a key objective being to measure the water surface profiles associated with ice jam formation, or the propagating waves associated with ice jam release (Kowalczyk and Hicks 2003). A key variable to river ice studies is streamflow, as the variable ice conditions occurring in the freeze-up, winter, and breakup periods pose a challenge to conventional measuring techniques.

3.2 Discharge Determination under Ice Affected Conditions

Currently the only reliable method for determining discharge under ice-affected conditions is to conduct direct measurements. This involves the use of a current meter to obtain point velocity measurements at (typically) two points in the flow depth, at more than 20 vertical panels across a channel. These point measurements are then integrated over the flow area to determine the total discharge. Pelletier (1989) provides a detailed description of typical practices for streamflow gauging under ice-affected conditions in both Canada and the USA.

Because of the cost and logistical difficulties associated with direct measurement, winter discharge estimates may be inferred from as few as two direct measurements over a six-month winter period (Moore et al. 2002). Indirect determination of the streamflow is a relatively straightforward procedure under open water conditions. Data collected with automated water level recorders can be readily converted to discharge using established stage-discharge relationships (i.e., rating curves), developed by conducting simultaneous water level and direct discharge measurements over a wide range of stream flows. Whenever possible, gauging sites are placed in reaches relatively unaffected by backwater and drawdown, so that there is little scatter in the stage-discharge relationship. When this is achieved, the stage-discharge relationship is quite adequately defined by a simple uniform flow approximation (e.g., Manning's equation).

Although the stage-discharge relationship is a well researched problem in open channel flow situations, and is even amenable to prediction through hydraulic analyses for highly dynamic flow situations (where a looped rating curve occurs), at present this relationship is poorly defined under ice conditions. However, it is known to be highly variable, particularly when partial ice covers and/or ice jams are involved. The primary reason for this is that, under ice conditions, it is not possible to develop a unique relationship between stage and discharge, as variations in ice conditions result in a non-unique rating curve. Despite this variability, it is possible to quantify the relationship between stage and discharge at a given river section under simple ice-covered conditions, providing ice characteristics are known and gradually varied flow hydraulics are considered (Hicks and Healy 2003). Nevertheless, direct discharge measurement remains the principal method in determining winter discharge.

Recently, there has been interest in developing faster and more efficient methods for conducting direct discharge measurements in winter, necessary not only due to cost and access, but also because of the hazards to the operational staff. It is difficult to determine when the ice cover is safe, and it will become a greater concern if climate change results in thinner and more intermittent ice covers. Water Survey of Canada (WSC) staff began researching the suitability and accuracy of ultrasonic (acoustic) flow metering devices in the early 1970s, and have been pursuing the goal of automated discharge measurement using these devices in earnest since 1985 (Wiebe et al. 1993). The advantage of acoustic flow metering is that it can potentially eliminate the need for streamflow gauging using conventional current metering techniques, replacing it with an automated measuring system. Healy and Hicks (2004) explored the applicability of empirical index velocity measurements which have the potential to speed up the measurement procedure, by providing accurate relationships between a single point velocity (or single vertical velocity profile) and the channel mean velocity. Morse (2005) has been investigating analytical relationships that have the potential to expand the applicability of this approach.

3.3 Remote Sensing Techniques

The use of Synthetic Aperture Radar (SAR) imagery as acquired by the Canadian RADARSAT and other satellites for sea ice monitoring has reached the operational stage and indications are that it has similar potential in the field of river ice monitoring. Weber et al. (2003) in a study on the Peace River, employed a visual analysis combined with a semi-automated classification system to categorize seven different "ice types" ranging from open water and frazil pan ice covers to several levels of juxtaposed and consolidated ice covers. Further research has revealed that depending upon the scale of the river and the resolution of the satellite image, it may also be possible to identify the extent of ice in a navigable waterway. For example, a variant texture analysis procedure has been successfully applied to both the Mississippi and Missouri Rivers to delineate specific regions of brash ice, sheet ice and open water on satellite images (Tracy and Daly 2003).

As part of the MAGS study, Hicks et al. (2006) investigated the potential for characterizing river ice using satellite SAR. Field data collected on the Athabasca River at Fort McMurray, Alberta, in winter 2003 lends credence to the suggestion that it may be possible to infer, indirectly, information on relative ice thickness from RADARSAT-1 imagery. In addition, heavily consolidated ice covers were observed to exhibit a higher backscatter than less consolidated adjacent areas. During the 2003 breakup period it was possible to distinguish areas of open water, intact ice, ice jams and running ice in three satellite images taken from April 18 to April 22 (Fig. 9 in Woo and Rouse 2007). This shows promise as a cost-effective means of monitoring river breakup remotely. Further research investigating the protocols to be used for image analysis is underway, and this will aid in the analysis of images taken with different modes and incidence angles. In addition, further work investigating the potential for river ice characterization in freeze-up and winter is also underway.

4 Ice Jam Flood Forecasting

4.1 Overview of Ice Jam Formation and Release

While ice jams are formed during freeze-up, it is the breakup jams that have by far the greater flood damage potential, owing to the much higher flow that accompany the breakup events. Breakup jams are initiated where moving ice blocks encounter a river reach where the winter ice cover has not as yet been set in motion. Jam formation is thus linked to the mechanism of dislodgment of the winter ice cover and the onset of breakup (Beltaos 1997). In natural streams, this mechanism involves channel planform, bathymetry, and slope, as well as ice thickness and strength. Sharp bends, sudden reductions in slope, constrictions, or islands are frequent jamming sites, along with areas where the ice cover is relatively thick and strong.

Once primed, jams propagate upstream at a speed dictated by ice supply, flow and channel conditions, as well as internal strength of the accumulated ice blocks that comprise the rubble. During this time, local flow and ice conditions can be highly dynamic. Eventually, an approximation to a steady state may be established so that jam and flow properties change little with time. Given sufficient supply of ice blocks, a jam may be many kilometers long, and attain its full potential thickness and water level, becoming an "equilibrium" jam (Fig. 3). This appellation derives from the development of a reach with relatively constant flow depth and thickness, termed the equilibrium thickness (Uzuner and Kennedy 1976). Downstream of the equilibrium reach, the water surface slope increases rapidly so that the water level profile can meet the much lower stage that prevails at the toe (downstream end). Here, the thickness of the jam increases, and often leads to "grounding", particularly in steep and wide reaches (Beltaos 1995).

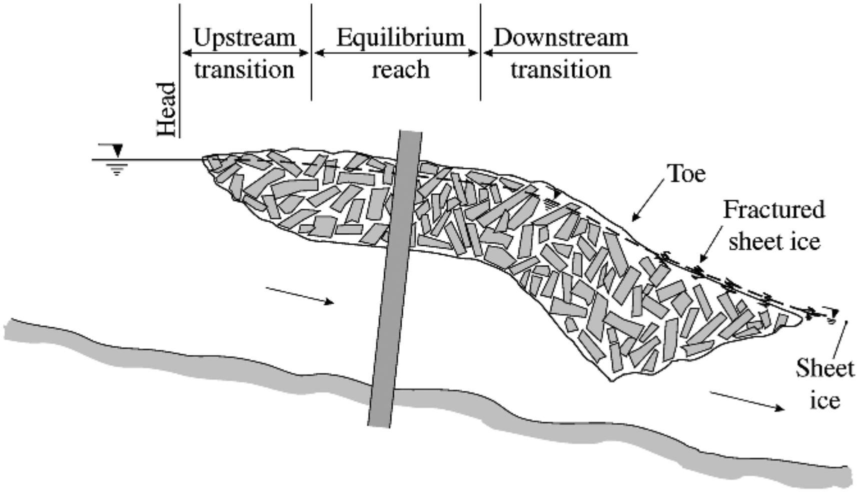

Fig. 3. Schematic illustration of an equilibrium jam. (From Beltaos (2001) with permission from ASCE)

Increasing flow and advancing thermal decay at the toe of an ice jam eventually cause it to release. Usually, the release is an abrupt event that generates a steep wave, as a large volume of water upstream of the toe can now move without obstruction other than bed friction. Release waves are characterized by rapid water-level rise (up to 0.8 m min^{-1} has been measured; Beltaos and Burrell 2005a; Kowalczyk and Hicks 2003), and very high water and ice velocities (several meters per second is not uncommon). Such waves have great damage potential, including the loss of human life. As already mentioned, release waves can also cause breakup of the winter ice cover over extended reaches, and are the main breakup-driving mechanism in flat rivers.

As part of the MAGS study, the hydro-climatic conditions leading to major ice-jam flooding of the Peace-Athabasca Delta have been identified (Beltaos et al. 2006a). Such flooding is essential for the replenishment of Delta lakes and ponds, especially those situated at higher elevations (perched basins). It was noted that significant jamming can only occur during a mechanical breakup event, and methodology to quantify the threshold between mechanical and thermal breakup was developed (Beltaos 2003a).

4.2 Empirical, Analytical, and Modeling Methods

It is not yet possible to predict the time and location of an ice jam, aside from likelihood statements in specific river reaches based on past experience. Ice-jam prediction is thus limited to thickness and attendant water levels, using analytical or numerical approaches. A key assumption is that the rubble in the jam behaves as a floating granular mass that obeys the Mohr-Coulomb failure criterion. The strength of the rubble is essentially supplied by the lateral confinement and by the effective upward stress, which is generated by the buoyancy of the ice in the rubble. When the jam forms, the rubble attains a thickness that is just enough to resist the internal stresses generated by the external forces (Pariset et al. 1966). Guided by this formulation, Beltaos (1983) obtained a simple functional relationship for equilibrium ice jams, and calibrated it with field data from Canadian rivers, spanning several orders of flow magnitude. The water depth within the equilibrium reach of a jam (Fig. 3) is primarily determined by river discharge, width, and slope.

The equilibrium water depth is a conservative estimate, because many jams do not attain the equilibrium condition. This is usually caused by limited supply of ice blocks and can be an important design factor in many practical situations. Steady-state, non-equilibrium jams are adequately predicted via one-dimensional numerical models, such as ICEJAM (Flato and Gerard 1986), RIVJAM (Beltaos 1993), and HEC-RAS (http://www.hec.usace.army.mil/software/hec-ras/hecras-download.html). These are public-domain models, but there are also proprietary models, such as ICEPRO and ICESIM (Carson et al. 2001). The proprietary DYNARICE model is both dynamic and two-dimensional (Liu and Shen 2000), and has been applied to field conditions. A radically different type of model is the DEM (discrete element model), which does not need to invoke the concept of a granular continuum. Instead, the motion of each block within the jam during small time steps is predicted by computing the forces applied on each block by the water and by the surrounding blocks. This approach provides important insights as to both the development and the final configuration of an ice jam and enables prediction of the forces exerted by jams on structures (Daly and Hopkins 1998; Hopkins and Tuhkuri 1999). As part of the MAGS study, the model RIVJAM was applied to the lower Peace River to determine the river discharge that is required for flooding of the Peace-Athabasca Delta when ice jams are present (Beltaos 2003b).

An important question pertaining to ice-jam flooding applications is how to determine the design stage for a variety of structures, including homes in residential subdivisions. For instance, when designing a bridge, the superstructure must be placed high enough to minimize the probability that it will find itself in the path of moving or jammed ice during the life of the structure. Such practical considerations underline the need for rational design criteria, based on stage-frequency relationships for ice-influenced conditions. Such relationships can be developed from historical records (empirical) or synthesized from local conditions (analytical).

4.2.1 Empirical Flood Frequency Estimates

The peak stages that can occur during breakup are strongly site-specific. Therefore, existing data should pertain to the site of interest or to its immediate vicinity. Transpositions and extrapolations should, as a rule, be avoided. Historical water level data may be available from various sources (e.g., hydrometric gauging stations, local residents, archives, photos, etc.) Ice scars on nearby trees also provide an indication of stages that occurred in the past and the year of occurrence (Gerard 1981). A method for performing a probability analysis on data deriving from such diverse sources as above is described by Gerard and Karpuk (1979). It is based on the "perception stage" concept (i.e., the stage below which any particular source would not have perceived, and recorded, the peak water level).

4.2.2 Analytical Flood Frequency Estimates

Crude indications of ice-caused flood frequencies can be obtained by a simple analysis based on the frequencies of the flow magnitudes occurring during the breakup period. A lower bound is obtained for the situation where no jams form near the site of interest. Then the peak stage can be calculated as a function of discharge using estimated values of ice cover thickness and hydraulic roughness. Similarly an upper bound may be established using the equilibrium-jam stage or by applying numerical models. For both the lower and upper bounds, the frequency of a given stage is that of the discharge associated with that stage. The actual frequency distribution will be somewhere between the two bounding distributions, and can be calculated if the probability of ice jam formation, P(J), near the site of interest is known (Gerard and Calkins 1984). For high flows, the ice-jam stage may be limited by the configuration and elevation of the river floodplain, while further flow increases may result in jam release ("ice-clearing" flow) and open-water stages (Tuthill et al. 1996). In general, P(J) could decrease with increasing flow (Grover et al. 1999), becoming zero when the flow exceeds the ice-clearing value.

4.3 Hydraulic Methods: Release Wave Modeling

Because of the dynamic nature of river ice jam release events and the significant flood risk they pose, it is desirable to be able to predict the speed and magnitude of the resulting release waves. This is a highly complex problem involving not only dynamic flow hydrodynamics but the interaction of the ice and water. Attempts have been made to model the propagation of ice jam release waves, most using one-dimensional hydrodynamic models and neglecting ice effects on the propagating wave (e.g., Blackburn and Hicks 2003). Although reasonable approximations could be achieved, difficulties were encountered in fully matching the shapes of measured stage hydrographs, suggesting that ice effects cannot be neglected. Jasek (2003) conducted field investigations documenting ice jam release events and found that the release wave celerity was affected by different ice conditions. Liu and Shen (2004) further explored this issue by applying a two-dimensional coupled flow and ice dynamic model (DynaRICE) to investigate the ice resistance effects (both internal resistance and boundary friction resistance) on ice jam release wave propagation in an idealized channel. Comparisons between the simulation results obtained with and without inclusion of ice dynamics showed that the ice effects reduce the peak discharge and slow down the release processes.

As part of the MAGS study, a number of ice jam release waves were measured on the Athabasca River at Fort McMurray. The most significant of these was in 2002 (Kowalczyk and Hicks 2003) during which the water level immediately downstream of the releasing ice jam rose 4.4 m in just 15 minutes. She and Hicks (2005) successfully modeled this event incorporating an uncoupled ice mass conservation equation and empirical momentum effects in the River 1D hydrodynamic model.

4.4 Regression and Logic-Based Ice Jam Forecasting Methods

Research into forecasting river breakup began over 50 years ago with simple single variable models and quickly led to multivariate models. Early work in this field was led by Shulyakovskii (1963) who viewed river breakup to be the result of deteriorating ice thickness and increases in river flows. Although Shulyakovskii's theory of river break up was simplistic, he did recognize the importance of the energy cycle in river ice processes and atmospheric circulation. Shulyakovskii quantified the concept of forecasting maximum stage rise during river break up as a function of several variables, including: ice thickness in the contributing reach; snow depth on the ice cover prior to ice melt; change in water level between the commencement of snow melt and the time of ice jam formation; the onset of negative air temperature in the ice-break up period; and total heat input. Both threshold models and regression models have evolved from these basic functional relationships.

Multivariate threshold models have been applied with modest success (e.g., Galbraith 1981; Wuebben and Gagnon 2005). However, specific additional variables and weighting factors required made these models highly site specific. White and Daly (2002) used stepwise selection of meteorological and hydrologic parameters to identify statistically significant input variables and then applied discriminant function analysis to predict ice jam occurrence. Massie et al. (2001) developed an artificial neural network model to produce a daily forecast of jam/no jam that required 22 input variables. Probably the most significant limitation of all of these models is that, while they do provide an assessment of the potential for ice jam occurrence, they cannot provide a prediction of the anticipated flood levels that might accompany an ice jam occurrence. Furthermore, they all exhibit a high degree of false positive indications of ice jam occurrence.

Single and multiple regression analyses have been applied to the problem of breakup water level prediction with moderate success. As part of the MAGS study, Robichaud (2003) developed a short term multiple linear regression model for predicting the maximum breakup water level for the Athabasca River at Fort McMurray, based on late winter snow pack and ice thickness as well as meteorological conditions in the few days immediately before breakup. Mahabir et al. (2007) extended the forecast lead time for Fort McMurray by several weeks, employing Fuzzy Expert Systems, a modeling technique based on set theory that allows variables to be described in linguistic terms. Fuzzy Logic has been applied successfully in a variety of fields where the relationships between cause and effect (variables and results) are difficult to express numerically but are conceptually well defined. Fuzzy Expert Systems produce a result based on logical linguistic rules rather than historical data, which allows this type of modeling to be less dependent upon the volume of historical data than many statistical methods. For the MAGS study, using only antecedent basin moisture, late winter snowpack conditions, and late winter ice thickness data, Mahabir et al. (2007) developed a Fuzzy Expert System that can forecast the potential for high water levels at breakup. Forecasts based on conditions known on April 1 (typically 3 weeks in advance of breakup) identified five years (of 22) for potentially high water levels at breakup, including all four actual high water events.

5 Climate Change Impacts on River Ice Processes

Numerous studies have examined the effects of climate on the timing of the ice season, using readily available data from hydrometric and related archives. It has been found that the general warming experienced in northern parts of the globe during the last 50 to 100 years is in step with a reduction in the duration of the winter ice cover and earlier occurrence of breakup on rivers (e.g., see review articles by Beltaos and Burrell 2003; Beltaos and Prowse 2001; Prowse and Beltaos 2002). The converse is true in isolated instances where the change involves cooling (Brimley and Freeman 1997). Andrishak and Hicks (2007) investigated the potential impacts of climate change on the thermal regime of the Peace River, as part of the MAGS study.

Such findings do not address the question of how climate change may alter the frequency and severity of extreme ice jam events. Relevant data are not as easy to find as freeze-up and breakup times, and their interpretation may be complicated by local precipitation and hydrograph characteristics. Consequently, prediction of breakup-related events would have to be based on a case-by-case analysis of current conditions and output from GCM/RCM-generated scenarios. A wide-ranging change that can be anticipated at present is the increased incidence of mid-winter breakups in parts of Atlantic Canada, Quebec, Ontario and British Columbia. This trend is already evident in parts of Atlantic Canada (Beltaos 2002, 2004; Beltaos et al. 2003). Mid-winter breakups are also expected to appear in certain regions that do not presently experience such events, such as the Prairies and northern Ontario and Quebec. For North America, Prowse and Bonsal (2004) applied simple but quantitative criteria to delineate present and future zones where mid-winter breakup events are likely to occur ("temperate region"; Fig. 4).

In certain rivers, mid-winter thaws can deplete the snowpack that is available for spring runoff, and thence reduce the severity of ice jamming during the spring breakup. This eventuality has been identified for the Peace River, as part of the MAGS study. Under future climate scenarios (2080s), depleted snowpacks are expected to severely inhibit ice-jam flooding in the lower Peace River, a necessary agent of replenishment of the lakes and ponds of the Peace-Athabasca Delta (Beltaos et al. 2006b, 2007).

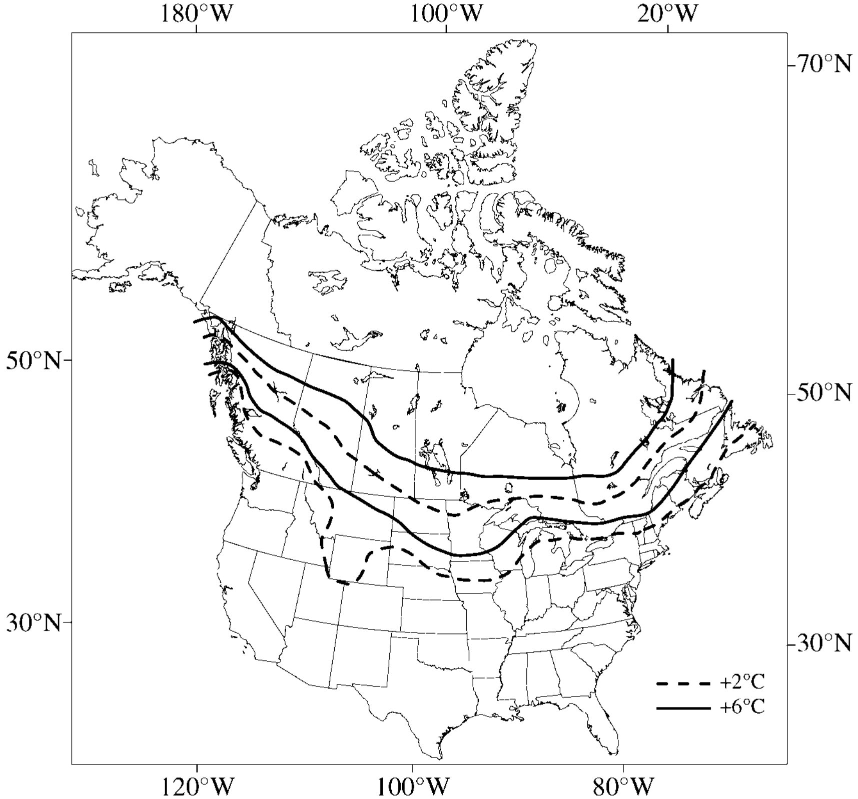

Fig. 4. Projected shift in the temperate region of North America based on mean winter temperature increases of 2 and 6°C. The shaded area is the current zone. (From Prowse and Bonsal (2004) with permission of the authors and of Environment Canada)

6 Summary

River ice plays a significant role in the lives of Canadians and other residents of high latitudes, providing a means of transport in the north, and threatening many communities across the country with flooding on an annual basis. River ice also plays a fundamental role in basin hydrology, and was a significant research component of the MAGS program. Investigations resulted in new methods for measuring streamflow under ice and the development of satellite remote sensing techniques for river ice characterization during breakup. Novel methods were developed for ice jam flood forecasting and to predict the flood waves associated with ice jam release. New, public domain, numerical models of river ice processes provide the means to assess the effects of climate change on the winter regime of rivers. Through a decade of field and modeling research, MAGS has substantially advanced the knowledge on river ice in cold regions.

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