Effects of climatic variability and flow regulation on ice-jam flooding of a northern delta

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Abstract:
Ice-induced backwater has been shown to be the only method by which flooding has supplied water to perched basins within the Peace–Athabasca Delta, one of the world’s largest freshwater deltas. The frequency of such events, however, markedly declined in the mid-1970s. To explain this shift, various hydrometeorological conditions that control the severity of river ice break-up were analysed. Specific emphasis was placed on the roles of flow regulation and climate variability. Flow regulation seems to have produced only minor changes in factors such as ice thickness and strength, and not to have reduced the flow at the time of break-up. Moreover, regulation has actually led to an increase in spring flow originating from the headwater region. Since the mid-1970s, however, spring runoff has declined in the downstream portions of the basin unaffected by regulation. This has been linked to a decrease in the magnitude of the winter snowpack. Elevated ice levels and winter flows resulting from regulation have further reduced the potential of tributary runoff to produce severe break-up floods. Thus the absence of a high-order event between 1974 and 1992 seems to be related to a combined effect of flow regulation and the vagaries of climate.

KEY WORDS climate variability; ice-jam flooding; flow regulation; Peace–Athabasca Delta

INTRODUCTION
Flooding is critical to the ecosystem health of river delta environments, particularly to perched ponds and lakes hydraulically separated from the open water channel system. One of the world’s largest deltas, the Peace–Athabasca Delta (PAD) in western Canada (Figure 1), has experienced two major periods of drying over the past three decades. These events have produced a range of changes in the delta ecosystem, varying from shifts in dominant vegetation regimes (Jaques, 1990) to dramatic declines in small mammal habitat and population (e.g. Townsend, 1975). The first drying trend was produced by the construction and rapid initial filling of the W. A. C. Bennett hydroelectric dam in the headwaters of the Peace River from 1968 to 1971. The second drying trend began after 1974, the year of a major spring flood, and affected the higher elevation landscape of the delta region, especially an area of isolated perched basins near the Peace River.

Although it was originally believed that decreased flooding of many of the high elevation perched basins near the Peace River was a result of reduced summer flows resulting from regulation, Prowse and Lalonde (1996) showed that ice-induced backwater was the historical source of annual high water levels. Peak levels for ice-affected events, before and after regulation, exceeded by as much as 2 m an historically high open

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Figure 1. Location of Peace River basin and the Peace–Athabasca Delta
The objective of this study was to determine if changes have occurred in the major hydrometeorological conditions controlling break-up that might explain the dramatic contrast in the frequency of large break-up floods before and after regulation. In a preliminary assessment of flow at break-up, Prowse and Lalonde (1996) suggested that break-up severity was related partly to variations in tributary flow downstream of the point of regulation, and that such variability might be controlled by the size of the accumulated winter snowpack. They also noted that other factors, such as ice thickness and strength, might be important. This paper evaluates the major hydrometeorological conditions that affect the resisting (thickness and mechanical strength of the ice cover) and driving (nature and origin of flow contributions) forces that have produced break-up and ascribes their temporal variations to the effects of flow regulation and climate. Specific details of some heat flux models are contained in the appendices and further details can be found in Prowse et al. (1996a).

SITE DESCRIPTION

The Peace–Athabasca Delta (PAD) is located adjacent to the lowest portions of the Peace River, approximately 1100 km below the point of regulation near Hudson’s Hope, British Columbia, Canada (Figure 1). The total drainage area of the Peace River (referenced to the Peace Point hydrometric station) is 293 000 km². The PAD is formed by the Peace, Athabasca and Birch rivers at the western end of Lake Athabasca in the province of Alberta (Figure 1). Large lakes occupy the central portions of the low relief, 3900 km² delta and

Figure 2. Open water rating curve for Peace Point hydrometric station and peak water levels produced by ice-jams. Dotted line indicates historical maximum water level achieved under open water conditions.
are connected to the major rivers by a series of channels. Water levels on the connected lake and channel network are typically highest during the spring and summer but then recede during late autumn and winter. Upstream flow regulation of the Peace River reduced summertime peak flows and generated higher winter flows, resulting in a net decline in delta lake water levels (Peace Athabasca Delta Working Group, 1973). In response, weirs were installed within the PAD to regulate water levels by restricting the northward flow of water and these have been largely successful in restoring summer peak water levels (Prowse et al., 1996b).

During periods of high lake and river water levels, water can also inundate the adjacent landscape and fill the myriad of shallow perched basins. Filling of the basins occurs either via a reconnection of abandoned channels or as a result of overland flow produced by overtopping of river banks and levees. In the case of the large zone of perched basins adjacent to the Peace River, Prowse and Lalonde (1996) have shown that backwater associated with ice break-up is the primary method by which such overtopping occurs.

Following the retreat of flood waters, decreases in the water levels of these isolated perched basins occurs almost exclusively by evapotranspiration (Prowse et al., 1996c). Long-term averages in precipitation and open water evaporation are approximately 450 and 380 mm, respectively, resulting in an annual water deficit of 70 mm. Given that groundwater flow through the levees is negligible (Nielsen, 1972), periodic flooding of these perched basins is essential for their ecological sustainability, both in terms of the aquatic habitat they provide and the various waterfowl and mammals that use them (e.g. Townsend, 1984).

HYDROMETEOROLOGICAL CONDITIONS CONTROLLING BREAK-UP

The greatest variations in flow and water level occur under dynamic break-up conditions and generally increase with river discharge and thickness and competence of the ice cover (e.g. Ferrick et al., 1992). Although maximum break-up water levels are known to occur within equilibrium reaches of ice-jams (e.g. Beltaos, 1995), insufficient data are available for the PAD area to conduct such an analysis of past events. Hence, the previous analysis of Prowse and Lalonde (1996) focused only on interannual comparisons of break-up water levels, based on the reasonable assumption that they vary with break-up severity.

Hydrometeorological conditions characterizing the most severe form of dynamic break-up usually include the generation of a large spring floodwave produced by the rapid melt of a large winter snowpack. Such conditions offer little possibility for the thermal decay of the river ice cover. On northward flowing rivers (warm to cold climatic gradient; e.g. see Gray and Prowse, 1993) where break-up advances in a downstream direction, the floodwave typically pushes into a reasonably competent ice sheet, all the while creating backwater behind the advancing, thick break-up front that periodically surges and stalls with the formation and release of ice-jams. On the Peace River near the PAD, ice-jams develop primarily from ice on the main stem with only minor contributions from tributaries.

The other extreme of break-up is a thermal event, one in which hydrometeorological conditions include low spring runoff, usually the result of a small winter snowpack or protracted melt, and extensive decay of ice thickness and strength. Ultimately, the remnant ice cover is so thermally weakened it can be dislodged by discharge comparable to a low flow winter period. Only minor increases in water level result from such events and formation of ice-jams is uncommon.

In general, river ice break-up severity is controlled by a balancing of the forces imparted by the upstream ice and flow conditions and those operating to keep the cover in place and intact, i.e. the resistance forces. Changes to either could explain why there appear to have been temporal changes in break-up events affecting the Peace–Athabasca Delta.

Resistance forces

The resistance of a river ice cover to break-up is principally determined by its thickness, mechanical strength and attachment to the bed and banks, although bonding is a relatively minor factor once hinge cracks develop. Unfortunately, except for ice thickness, no regular measurements of these variables are undertaken. Since changes to all these resistance factors are controlled by atmospheric and hydrothermal
heat fluxes, interannual variations of their condition can be estimated through modelling of meteorological heat exchanges and approximations of flow-related heat fluxes.

Winter ice growth. One measure of ice-sheet resistance is the thickness of the cover at the time of break-up. Interannual variations in thickness could result from changes in hydrometeorological conditions produced by climate and/or flow conditions. In the case of a regulated system, a number of factors may affect ice growth and final thickness. These include changes in ice cover duration, initial ice thickness and hydrothermal heat fluxes.

(i) Timing and duration of Peace River ice regime. In a companion study, Conly and Prowse (1998) analysed historical records of ice conditions along the Peace River and showed that flow regulation significantly altered the timing and duration of the ice regime upstream of the town of Peace River. Close to the dam, the ice season has been virtually eliminated. Further downstream, only an intermittent ice cover develops and, at the town of Peace River, there has been a significant delay in the initiation of freeze-up and the overall ice season. At the downstream extremity of the Peace River near the PAD, however, regulation has not significantly affected the timing or duration of the main ice season and, therefore, the conditions for ice growth.

(ii) Initial freeze-up thickness. The winter thickness of an ice cover can also be strongly influenced by its initial freeze-up thickness, as dictated by the prevailing hydraulic conditions (i.e. slope, velocity, flow depth, ice type and geometry). In general, a thick initial accumulation of porous frazil ice leads to a greater total solid ice thickness because of its enhanced freezing potential (e.g. Calkins, 1979). Although higher velocities associated with increased regulated flow during freeze-up are likely to produce an enhanced initial cover thickness (consolidated frazil ice floes) in the steeper upstream reaches of the Peace River, this is not the case in the lower slope reaches that characterize the reaches near the delta, where the initial cover develops from the juxtapositioning of ice floes (Andres, 1996).

(iii) Hydrothermal fluxes. Ice thickness can also be significantly affected by the introduction of warm water to a river system. Inputs of warm hypolimnetic water from the Williston Reservoir are responsible for the delay in freeze-up of the ice cover in the upstream reaches of the Peace River (Andres, 1996) and are probably also responsible for some retardation of ice growth near the advancing freeze-up front. Under the turbulent flow conditions and large ice cover roughness experienced on the Peace River, however, water temperatures would be rapidly reduced to 0 °C within a relatively short distance (i.e. ~ < several hundred metres) of the freeze-up front (i.e. interface between the upstream open water and downstream ice cover). As a result, warm hypolimnetic water would not affect ice growth conditions in the lower river reaches of interest to this study. Although heat from groundwater is a direct contributor to the total hydrothermal flux, it was assumed to remain relatively constant from year to year in the source areas downstream of the reservoir.

Increases in flow can also mean an increase in heat flux to the ice cover because of increased fluid friction, $N_f$, as defined by

\[ N_f = \rho_w g S V h \]  

where $\rho_w$ is the density of water (kg/m$^3$); $g$ is the acceleration due to gravity (m/s$^2$); $S$ is the energy slope; $V$ is the mean flow velocity (m/s); and $h$ is the flow depth (m). The heat produced by fluid friction can either retard ice growth during the winter period, when the overall energy balance is negative (positive net flux to the atmosphere), or accelerate ice ablation during the spring, when the energy balance is positive (positive net flux from the atmosphere). A general understanding of the potential increases in fluid friction that might result from increased winter flow can be gained from Table I, which shows the average pre- and post-monthly flows for the Peace Point hydrometric station and the related heat flux associated with fluid friction.

Assuming a similar slope of 0.00011 and river width of 700 m (Hicks and McKay, 1996) for both scenarios, the enhanced flow would result in an approximate threefold increase in $N_f$, from an average of less than 1 W/m$^2$ for the pre-regulation flows to approximately 2.4 W/m$^2$ for the post-regulation winter period.
Such values are small, however, compared with the winter atmospheric heat fluxes promoting ice growth (see discussion below).

Although $N_f$ is a relatively small term in the energy budget, it can accumulate to a substantial total over the entire winter period. The total increase in the fluid friction heat flux over the four main months of winter ice growth (December to March) can be converted into a ‘melt equivalent’ of ice ($t_m; m$) from

$$t_m = \sum \phi_f / (\lambda_i \rho_i)$$

where $t_m$ = the total ‘melt equivalent’ of ice (m); $\sum \phi_f$ = the accumulated heat due to fluid friction (J/m$^2$); $\lambda_i$ = the latent heat of fusion for ice (J/kg); and $\rho_i$ = the density of ice (kg/m$^3$).

Assuming all such heat is transferred to the overlying ice cover, the result would be a melt equivalent of approximately 56 mm for the four-month winter period. The significance of this to overall peak ice thickness on the Peace River is discussed below.

(iv) Observed peak ice thickness. To evaluate further, whether there has been any temporal trend in late season ice thickness for Peace Point, data were compiled from the Water Survey of Canada (WSC) records from the Peace Point hydrometric station. Only late season measurements were used but their timing varied and in some years no measurements were conducted several months prior to break-up. To obtain consistency in the data for comparative purposes, only those measurements conducted between 14 March and the beginning of the spring melt period were used. Determination of this latter date is described in the subsequent section, which focuses on ice ablation. When more than one measurement survey was conducted in a given year within this time frame, the maximum value was used. The pre-break-up peak ice thickness averaged 0.86 m (standard deviation, $\sigma = 0.08$; number of cases, $n = 9$) while that for the post-regulation period was 3 cm greater at 0.89 m ($\sigma = 0.13$; $n = 15$). The difference is not statistically significant (significance level, $\alpha = 0.05$), although this comparison of measured data is affected by a number of factors. These include a one-month span in the end of season data (i.e. 14 March to 16 April, although growth rates are expected to be small at the end of the season) and interannual variations in: (a) the location of the ice surveys; (b) the concentration of frazil ice — known to alter ice growth rates (e.g. Calkins, 1979); and (c) snow load — which directly affects the WSC determination of ice thickness (e.g. Adams and Prowse, 1986).

(v) Modelled peak ice thickness. Given the potential errors in the above comparison of measured data, peak ice thicknesses were also modelled using a degree–day approach (Michel, 1971). Although inaccurate in the early stages of ice growth, this approach has been proven to provide reliable results for established covers when varied to account for different growth environments. Thickness at any given time can be estimated according to

$$t_i = \kappa D_t^{0.5}$$

where $t_i$ = ice thickness (mm); $\kappa$ = a coefficient varied to account for conditions of exposure and surface insulation (mm/°C$^{1/2}$ d$^{1/2}$), and $D_t$ = accumulated degree–days above freezing (°C d).


<table>
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<th>Melt (mm/month) Pre</th>
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To test the applicability of this approach, all ice thickness measurements were assembled from the WSC records and plotted according to their respective value of $D_f$. Degree–days were accumulated from the date of freeze-up, as determined from the original hydrometric records for each year. Air temperature data were obtained from the national climate station at Fort Chipewyan. The results are shown in Figure 3a. Statistically, the most suitable value of $k$ for the Peace Point data was found to be 0.18 with an $r^2$ of 0.61, which corresponds closely to the range of 0.14 to 0.17 suggested by Michel (1971) to be representative of conditions on an ‘average river with snow’. Unfortunately, the lack of air temperature data prior to 1963 resulted in a short pre-regulation record of modelled ice thickness values.

In modelling ice thickness, the date of ‘peak ice thickness’ was assumed to be the date of the first significant spring melt as described in the subsequent section on ice ablation. The results are shown in Figure 3b. The mean peak thicknesses were 0.96 m ($\sigma = 0.033; n = 5$) for the period 1963–1967 (pre-regulation) and 0.94 m ($\sigma = 0.061; n = 21$) for the longer post-regulation period (1972–1992). As was the case for the measured
values, these two means are not statistically different, although the data suggest there may be some form of decreasing trend with time (Figure 3b). Notably, however, this does not correspond with the time of regulation but begins to occur in the mid-1970s, possibly indicating some shift in climatic conditions.

**Spring ice ablation.** The thickness of the pre-break-up ice cover and its associated attachment to the banks and bed can decrease because of atmospheric and/or hydrothermal melt. Unless there is a significant input of warm water from groundwater or tributary flow, bottom melt of an intact ice cover by the hydrothermal flux is relatively small. The main ablation results from surface melt, first of the snowpack and then of the underlying ice sheet. Unfortunately, few data are ever collected about rates of river ice ablation because of the logistical and safety problems associated with obtaining such information. To determine whether there exists any specific temporal trend in the seasonal ice ablation record, the controlling heat fluxes and pre-break-up ice thicknesses were modelled for the period 1959–1992. A detailed energy balance and a simplified degree–day approach were used depending on the availability and type of meteorological data. Details of each approach and descriptions of terms are noted in Appendix A.

(i) **Results of energy balance ice ablation calculations.** Summaries of the major atmospheric heat fluxes appear in Figure 4a for the pre-break-up melt periods of each year from 1963–1979 (i.e. the years in which hourly data were available to model the various fluxes). Although this record length is insufficient to analyse for significant changes in the time-series of the heat components, it does contain a good combination of break-up types. Specifically, it contains six of the seven break-up events that produced water levels exceeding that resulting from the historical open water, high flow event of 1990. It also includes numerous lower order events that occurred in the intervening years of the 1960s and 1970s. The following discussion reviews the heat fluxes that tend to dominate break-up and evaluates whether they are related to the severity of break-up events (as indexed by water levels).

The relative net heat flux values for each year are presented in Figure 4b. As shown, most events have been primarily high radiation melt events. On average, net radiation accounted for approximately 70% of the total atmospheric heat flux during the pre-break-up melt periods. As is to be expected for such clear-sky melt periods, the latent heat flux ($Q_e$) tended to be negative and contributed only to evaporation of the ice/snow mass, as opposed to additional melt through latent heat associated with condensation. Although the convective transport of sensible heat averaged only 25% of the net atmospheric heat flux ($Q_a$), it was significant in some years, often exceeding 20 MJ/m² and in 1964 was as high as 52 MJ/m² (Figure 4a). Overall, however, most melt periods were dominated by the radiative component. This was because of the dominance of the net short-wave component (Figure 4c) which exceeded the combined total heat flux in all years, including 1964 when significant contributions of sensible heat were added. As discussed, it is the solar radiation component that contributes not only to thickness decreases but also to internal melt and mechanical strength decreases. The implications of this are considered in a subsequent section.

As identified by Prowse and Lalonde (1996), the years 1963, 1965, 1967, 1972, 1974 and 1979 (those within the available record length for heat flux calculations) were characterized by significant (> the 1990 open water flood) break-up water levels (see Figure 2). The composition of the major heat fluxes associated with these years was compared with that for years of less severe break-up (i.e. as defined by lower water levels). On average, the large break-up years were characterized by a melt period of lower total $Q_a$ (average net $Q_a = 41$ MJ/m²) than those of other years (55 MJ/m²). This comparison excludes years of major reservoir filling (winter 1969–1971) when anomalously low flow conditions prevailed, atypical of both the pre- and post-regulation periods. For the non-fill years, 1972 was an exception with $Q_a = 61$ MJ/m². Part of the reason for the generally lower overall heat flux is that the period of melt tended to be shorter, averaging nine days (range = 7–15; $\sigma = 2.9$) for the large break-up years and 11 days (range = 6–19; $\sigma = 3.9$) for the other non-fill years. Such differences, however, are dependent on the degree of error involved with the original selection of the melt initiation date. Considering such differences in melt duration, the more severe years of break-up were actually characterized by a marginally lower rate of melt, 4.9 MJ/m²/d, compared with 5.7 MJ/m²/d.
Figure 4. (a) Modelled atmospheric heat fluxes during the pre-break-up ablation period. (b) Relative contributions of heat fluxes to net heat flux. (c) Composition of radiative heat flux.
In a further attempt to explore whether there was a relationship between break-up water levels and the magnitude of pre-break-up melt, $Q_a$ values were regressed against an ice-affected backwater level ($\Delta h$; Figure 5a). The latter was obtained by calculating the water level that would result from the discharge on the day of peak water level under open water conditions ($h_0$) and subtracting this value from the measured peak water level ($h_m$). Exclusion of the major reservoir-filling years results in $r^2 = 0.61$ between $Q_a$ and $\Delta h$. Given the complex factors affecting this relationship, and that $\Delta h$ values do not account for

Figure 5. (a) Net heat flux versus stage increase, 1963–1979. (b) Accumulated melting degree-days versus stage increase, 1963–1992. (c) Residual mass curve of accumulated melting degree-days
interannual spatial variability in ice jamming, the relationship is surprisingly strong. Moreover, the general nature of this relationship — decreasing backwater with increasing pre-break-up melt — makes practical sense in terms of the broad classification of break-up events defined earlier. Events plotting to the right-hand side of Figure 5a would fall into the category of thermal break-ups, i.e. ones associated with extensive melt (high heat input). Those to the left are typical of dynamic events when pre-break-up thermal inputs are at a minimum.

(ii) Results of degree–day ice ablation calculations. Although temperature-based indices are poorer indicators of melt conditions than energy balance approaches, the longer record of daily temperature data permitted calculation of $T_{\Sigma M}$ for a 30-year period (1963–1992) and a more thorough investigation of the relationship between heat input and break-up severity (i.e. $\Delta h$), and of long-term trends in pre-break-up melt conditions.

In the first series of degree–day calculations, attempts were made to define a suitable $\tau$ coefficient for use in equation (A2) (see Appendix A). Unfortunately, insufficient reliable data about spring decreases in ice thickness were available to establish a realistic value for $\tau$. Since, however, ice thickness is simply a linear function of $T_{\Sigma M}$, it was decided to perform the interannual comparison directly with the degree–day values. In calculating $T_{\Sigma M}$, a base reference temperature of $-5^\circ C$ was used because of its previous validation in ice ablation studies (e.g. Bilello, 1980; Prowse et al., 1989). Again, the lack of data before 1963 makes it difficult to compare pre and post-regulation conditions. Although there is no significant difference ($\alpha = 0.05$) between the $T_{\Sigma M}$ values associated with these two periods, there appears to be a tendency towards higher values after the mid-1970s. The major break-up events prior to regulation (1963, 1965 and 1967), and the three since regulation (1972, 1974 and 1979), occurred with $T_{\Sigma M}$ values less than 131, and no year prior to 1974 exceeded 134. In the period since, however, over half of the years equalled or exceeded this value. Such enhanced heating would lead to increased thinning of the ice cover and, all other factors being equal, to a lower ice thickness at the time of break-up. Notably, however, there have also been years during the 1980s with $T_{\Sigma M}$ values equal to or less than those associated with the large break-ups of the 1960s and early 1970s.

Regressing $T_{\Sigma M}$ with $\Delta h$ produced a correlation coefficient of only 0.37 (excluding filling years; Figure 5b). Although this is lower than that for the heat flux approach, the data again suggest that there may be a similar, albeit weaker, trend of decreasing $\Delta h$ with increasing heat input.

A longer duration record also permitted trend analysis of the degree–day data. To identify any potential temporal trend in the data, they were analysed by means of a residual mass curve, a method commonly used in the analysis of hydrometeorological data (e.g. Buishand, 1982). Unlike simple analyses, such as running means that smooth variability in temporal records, this method focuses more on identifying persistent trends in a time-series. The resultant time-series is plotted in Figure 5c and again suggests that there was a shift in the late 1970s (specifically 1979) to conditions characterized by generally higher heat input during the pre-break-up ablation period. The series was also tested for homogeneity using the von Neumann ratio, and $Q$ and $R$ values as outlined by Buishand (1982). To calculate the latter values, the cumulative deviations from the mean were rescaled first by dividing by the sample standard deviation. None of these tests suggested a strong shift in the mean within the time-series. A $t$-test for a shift in the mean at 1979 (Salas, 1993), however, (based on two time series on either side of this apparent break point) did indicate that there was a significant shift in the data at this point. Tests on other subseries within the record length showed no significant difference in subseries means.

If there is a shift to years of higher degree–days after 1979, the implications are increased melt prior to break-up and, if a relationship does exist between pre-break-up melt and break-up severity, a tendency to lower stage increases as a result of break-up. Aspects of this apparent shift in pre-break-up melt conditions are discussed more fully in a later section.

Changes to ice strength. Although ice strength is a critical factor in controlling the severity of break-up floods, no regular measurements of ice strength are made in hydrometric programmes. There are, however,
some basic relationships that relate ice strength to changes produced by internal radiative melt of ice covers. Details of the approach employed are outlined in Appendix B.

Results from the first set of calculations indicated extremely rapid changes in ice strength. Without direct field data, it was impossible to know whether these results were real or due to some inaccuracies in the current theoretical assumptions, such as the use of a bulk porosity/strength value for the entire cover. Since the objective of this section was to compare interannual differences in an index of ice cover strength at break-up, it was decided to employ a slightly different index approach, but one still based on the established exponential decay of strength with porosity. The second approach involved calculating radiation absorption and resultant porosity values for the middle 50% of the ice sheet and then the related strength index.

Neither a time-series nor a residual mass curve plotting of \( \theta / \theta_0 \) values, corresponding to the day of break-up for each year, exhibited any obvious temporal trend. Moreover, no clear relationship could be found with break-up severity (i.e. \( \Delta h \)), as might have been expected intuitively. Years of very high \( \Delta h \) were characterized by final strength ratios comparable with those for years of very low \( \Delta h \). Hence, based on the results of this approximation of ice strength, ice strength was not a significant factor in controlling variations in break-up conditions.

**Upstream forces**

**Flow contributions at break-up.** As noted above, the magnitude of backwater produced by break-up activity is a function of spring discharge. In a preliminary assessment of flows contributing to the day of peak break-up water levels, Prowse and Lalonde (1996) found that a downstream tributary, the Smoky River, contributed more flow than that originating from above the dam site, before and after impoundment. The current study furthers the analysis to include an evaluation of flow from other major, gauged downstream tributaries to both the initiation of break-up (\( h_i \) at Peace Point) and the subsequent peak break-up water level (\( h_m \)). Flow travel times used for lagging tributary flow (Table II) were derived from a hydraulic model of the river (Hicks and McKay, 1996). To minimize potential errors in travel times or precise definition of break-up dates, contributing flows were averaged over a three-day bracket, corresponding to the break-up dates.

In general, analysis of hydrometric records revealed a significant (\( z = 0.05 \)) increase in the average flow initiating break-up at Peace Point between the pre- and post-regulation periods (3050 m\(^3\)/s to 3418 m\(^3\)/s), but a decline in the flows that produced \( h_m \) (averaging 4128 m\(^3\)/s and 3771 m\(^3\)/s, respectively). Values for specific years are noted in Figures 6a and 6b. In terms of the sources of water that produced such flows, there are significant differences in above and below point-of-regulation and pre- to post-regulation values.

Although the pre-regulation record is relatively short, there is a striking contrast in flow conditions before and after impoundment. The average pre-regulation flows at Hudson’s Hope contributing to \( h_b \) and \( h_m \) were 565 and 1146 m\(^3\)/s, respectively, but these both increased by approximately 500 m\(^3\)/s with regulation. A similar doubling is evident in the relative percentage values for Hudsons Hope which increased from 19 to

<table>
<thead>
<tr>
<th>Drainage area (km(^2))</th>
<th>Percentage of drainage area (%)</th>
<th>Distance from Peace Point (km)</th>
<th>Estimated flood travel time (days)</th>
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</tr>
</tbody>
</table>

Figure 6. Actual and relative flow contributions recorded at main stem stations. (a)–(d) and major downstream tributaries (e)–(f) related to time of break-up initiation (left column figures) and peak break-up water level (right column).
36% and 17 to 33% for \( h_b \) and \( h_m \), respectively (Figures 6c and 6d). On average, therefore, it is clear that a greater proportion of the flow supplied to break-up near the PAD, both before and after regulation, was contributed by flow entering downstream of the dam site. The post-regulation average increase of 500 m\(^3\)/s can logically be attributed to the effects of flow enhancement by regulation for late-winter hydroelectric production. The change in percentage values, however, could also be a result of changes in the other contributing sources downstream of the reservoir, i.e. the major tributaries.

Flow records exist for five major tributaries between Hudson’s Hope and Peace Point, their combined catchment area representing 55% of the total contributing area between these two Peace River stations. Relevant data regarding drainage area, percentage of the total Peace River catchment area above Peace Point and approximate flow travel time to Peace Point are listed in Table II. In general, their response can be considered representative of the smaller downstream tributaries that drain the foothill and plains regions. Unfortunately, data for all tributaries were not available for the entire period of study, as noted below.

Based on a break-up contributing flow analysis (Figures 6e and 6f), it is clear that downstream tributary input at break-up is very much dominated by two river systems, the Smoky and Wabasca rivers. This is because of their large area and relative location in the catchment, which controls their runoff response. Together, the Wabasca and Smoky rivers comprise almost 40% of the catchment area between Hudson’s Hope and Peace Point. Moreover, much of this area is composed of foothills and lowland plains, zones that experience the earliest spring snowmelt. Virtually all of the Wabasca River lies below approximately 600 m and a significant portion of the Smoky lies below 900 m. Further upstream, the other three tributaries (Beaton, Halfway and Pine rivers) comprise only 16% of the area between Hudson’s Hope and the PAD but they also drain a greater proportion of higher elevation zones that experience later spring snowmelt. As a result, and as evident in Figure 6e and 6f, the amount of spring runoff from these types of tributaries contributing to break-up conditions near the PAD tends to be significantly smaller than that produced by the Smoky and Wabasca catchments. Detailed analysis of spring hydrographs for these catchments indicates that in most years their main spring contribution arrives after the major break-up has occurred at the PAD.

Unfortunately, of the two dominant tributaries, only the Smoky River has a pre-regulation record that can be used to explain the changes in relative flow percentages noted above. Even so, the dominance of even this one tributary relative to the zone upstream of the point of regulation is clear. For example, the average flow contributed to the 1962–1967 break-ups by the Smoky River was 736 m\(^3\)/s, or 24% (range 9–43%), an average 30% greater than that contributed by Hudson’s Hope. In terms of \( h_m \) values, the Smoky River contributed almost twice the flow measured at Hudson’s Hope, averaging 1183 m\(^3\)/s, or 28% (range 8–67%). In general, this one downstream tributary was on average slightly more important than the entire upstream flow in initiating break-up, and significantly more important in producing the peak break-up water levels. The larger importance of flow contributed from the downstream portion of the basin (below the point of regulation) compared with the upstream portion becomes more apparent if all the downstream tributary flow is combined, although the Smoky River still dominates. Note that the Wabasca River is not available for this pre-regulation period.

Since regulation, the average flow contributions of the Smoky River to \( h_b \) and \( h_m \) have significantly declined by approximately 40–50% from their pre-regulation values, to 468 m\(^3\)/s and 614 m\(^3\)/s, respectively. Such large reductions in downstream tributary flow offer a second reason for the large increase in the relative importance of flow from above Hudson’s Hope in contributing to break-up near the PAD. The reasons for such declines in tributary flow volume are considered in the subsequent section.

The above discussion has focused on average flow conditions. Importantly, however, the contrast in importance of upstream vs. downstream flow contributions is even stronger for years of major break-up. For example, considering all years of record, the average downstream contribution for lower order break-up events (defined as when \( h_m \) was less than that of the historically high 1990 open water flood event) was 65 and 68%, respectively, for \( h_b \) and \( h_m \), but much higher, at 74 and 79%, for the higher order events (i.e. 1963, 1965, 1967, 1972, 1974, 1979 and 1992). In the two years, 1965 and 1974, noted by Prowse and Lalonde (1996) to
have caused the most widespread documented flooding of the PAD, the downstream flow contributions to $h_m$ were 91 and 82%, respectively, of which 67 and 39% of the total flow originated from the Smoky River alone (65% for 1974 if the Wabasca River is included). Notably, the upstream flow contribution was high at break-up initiation for the 1974 event but declined to only 18% at the time of break-up water level. This dramatic change in percentage contribution was simply due to a rapid increase in the contributing flow from the downstream tributaries. The flow at Hudson’s Hope, in fact, remained relatively constant, rising from 1070 to 1180 m$^3$/s between the nine days separating $h_s$ and $h_m$.

In summary, the average relative contribution of the upstream flow above Hudson’s Hope has increased while the average contribution from the downstream tributaries (as indexed primarily by the Smoky River) has decreased. Some of the extreme events in both the pre- and post-regulation periods (e.g. 1965 and 1974), however, appear to be associated with abnormally high spring flows in the tributaries, especially the Smoky River. It appears, moreover, that tributary flow since 1974 has been insufficient to generate higher order break-up events.

**Snowmelt as the source of spring flow.** The relatively small flow contribution of the Peace River upstream of Hudson’s Hope during the pre-regulation period is consistent with the hypsometry of the Peace River Basin: the headwaters being largely within high elevation alpine regions of the Rocky Mountains. Snowmelt runoff from the higher elevation zones does not usually occur until June, well after break-up has occurred in the lower elevation downstream portions of the Peace River. The dominance of downstream tributary flow in controlling break-up flows on the lower Peace River is related to earlier snowmelt runoff produced by such tributaries as the Smoky and Wabasca rivers. Considering this, an analysis was conducted of snowpack records in an attempt to explain the shift in the mid-1970s to lower spring runoff from the downstream tributaries.

Fortunately, the station with the most complete record within the lower Peace system is located at Grand Prairie in the mid-elevation portions (668 m a.m.s.l.) of the Smoky River. No station exists in the Wabasca catchment. The water equivalent data for the maximum spring snowpacks for this station are shown in Figure 7a. All the major break-up events occurred with the peak snowpack (water equivalent) at Grand Prairie exceeding approximately 110 mm: the 1974 event also being the year of the long-term maximum peak snow water equivalent. Notably, other years of less severe break-up ($h_m$) were also characterized by large spring snowpacks (e.g. 1982). It should be remembered, however, that such snowpacks only indicate the potential for large spring snowmelt. Numerous other factors, such as the rate of melt and antecedent moisture conditions, control whether the large snowpack will produce a major runoff event. As noted by Prowse and Lalonde (1996), a year like 1974 was characterized by a pronounced above-normal spring runoff response in the Smoky River catchment: one most probably linked with intense melt of the above-normal snowpack. By contrast, analysis of the 1982 hydrograph revealed only a gradual rise in spring discharge resulting from a protracted melt of the above-normal snowpack. This reinforces the role of melt rate in controlling break-up severity.

Given the apparent importance of snow at the Grande Prairie station to runoff of the Smoky River, winter precipitation records were synthesized to produce a longer term record. Figure 7b shows the winter accumulated precipitation, from 1 November to 31 March, for each year of available record from 1947 to 1992. Also shown is a seven-year running mean through this data set. The snowpack record, like that of tributary runoff, indicates a general shift to below-normal values beginning in the mid-1970s; only the 1982 event is significantly above the long-term normal.

To explore further the potential existence of a trend in the snowpack records for Grande Prairie, the accumulated precipitation and peak snowpack data sets were converted to residual mass curves (Figure 7c). In both cases, there appears to be a downward trend in the data after 1974. This apparent mid-1970s shift in peak snowpack values has also been observed in adjacent British Columbia (e.g. Moore and McKendry, 1996) and has been shown, for the case of the Peace River catchment, to be related to changes in the frequencies of snow producing weather types (Keller et al., 1997).
Freeze-up effects. There is considerable evidence in support of a relationship between freeze-up stage, $h_f$, on a river and the stage at break-up initiation, $h_b$ (e.g. Shulyakovskii, 1966; Beltaos et al., 1990). Typically, break-up is not initiated until $h_f$ has been at least exceeded. As shown in Figure 8, there is a generally good relationship between $h_f$ and $h_b$ for the Peace Point hydrometric station. Data were extracted from copies of

Figure 7 (a) Annual maximum snowpack, Grande Prairie, Alberta. (b) Accumulated winter precipitation, November to March, Grande Prairie, Alberta. (c) Residual mass curves of snowpack and precipitation data from (a) (squares) and (b) (circles)
the original hydrometric charts following the methods outlined in Beltaos et al. (1990) and are shown categorized according to the phase of regulation. In general, the post-regulation data are skewed to higher freeze-up levels. Although data could not be extracted for some years, and despite the very short record prior to regulation, there has been a significant increase (1.2 m) in levels following regulation. Based on a smaller data set for this same site, Andres (1996) calculated a comparable value of 1.4 m. Such increases can be the result of two factors: increased autumn flows and greater staging potential. As noted earlier, the latter can be discounted because the relatively low slope reaches of the lower Peace River do not favour freeze-up formation by consolidation processes (thicker covers that promote increased freeze-up staging) even under highly elevated discharge values. Hence, the post regulation increases in $h_f$ are the direct result of higher flows. Based on monthly flow values, the ratio of average November flow (main month of freeze-over at Peace Point) for post- to pre-regulation conditions is 1.4. This ratio increases to an average of 3.3 for the main winter months, December to March.

Higher freeze-up levels may be related not only to break-up initiation, but also to overall break-up severity. Although an attempt was made to relate values of $h_f$ to a simple index of break-up severity, such as the maximum break-up water level, there was no obvious correlation ($r^2 = 0.16$). This is not unreasonable given the number and complexity of factors affecting this relationship. Despite the apparent difficulty in quantifying a relationship between freeze-up levels and break-up severity, there are some strong arguments that can be made as to why one should exist. The rationale for these arguments stems from two sources: (a) resiliency of elevated ice covers to ‘break-up’ discharge (e.g., Beltaos, 1997); and (b) the decreasing effect of tributary inflow with increasing main stem discharge.

First, considerable knowledge about the resiliency of ice covers to premature break-up has been gained through operating schemes employed by hydroelectric facilities during the freeze-up period. In general, as an ice cover becomes increasingly established (thickness and mechanical competency), flow can be increased significantly without causing the cover to break. In the case of the International Section of the St Lawrence River, for example, flows can be increased by up to 30% (Wigle et al., 1990, pp. 16 and 69). In simple terms, the higher a freeze-up cover is stabilized, the greater the flows it can withstand without breaking. Moreover, elevating an ice cover during the winter period through regulation of flow means that the river will be able to pass greater discharge in the spring without rupturing the ice cover. At the beginning of break-up, the main
stem discharge must exceed $h_f$ by some factor (e.g. 30%) to initiate any form of ‘dynamic’ break-up. Otherwise, the cover will simply continue to deteriorate in situ until it is so thermally weakened that a low magnitude flow is able to move it downstream, a situation more akin to a ‘thermal’ break-up and much less likely to produce significant backwater. Similar arguments about a relationship existing between $h_f$ and break-up severity have recently been made by Beltaos (1997). Extensive melt of an ice cover prior to break-up could also reduce the probability of large ice-jam formation by reducing the available ice supply, possibly even to the point where an equilibrium ice-jam could not develop.

Elevated ice levels produced by regulated flows do not mean, however, that the river will not produce large spring break-ups and associated flooding. If the spring flow significantly exceeds that associated with the elevated freeze-up stage, a severe break-up could still be produced. The amount by which spring flows exceed $h_f$ depends on the magnitude of flow supplied by the reservoir and the downstream tributaries. Although the regulated flow is typically higher at this time of year than would occur under unregulated conditions, further increases in regulated flow are unlikely because of the declining seasonal requirement for hydroelectric power. Moreover, the break-up months of April and May are typically the time that high winter flows begin to decline to lower summer values. For example, the ratio of post- to pre-regulation average monthly flows recorded at Hudson’s Hope decreases from 2.0 in April to 0.8 in May. Thus, to exceed the winter regulated flow (or $h_f$), significant contributions must come from tributaries downstream of the point of regulation. Furthermore, if the amount of regulated flow at the time of break-up is declining from higher winter values, tributary flow must also account for this ‘loss’ to the main stem discharge.

It is clear from the analysis of flow contributions at break-up that large tributary flows have led to the formation of large break-up floods affecting the PAD, both before and after regulation. For such major runoff events, it is unlikely that elevation of the winter ice cover would have much effect on the overall severity of the break-up event. Similarly, cover elevation is probably not important for years of very low runoff when the related small rise in main stem discharge would simply pass beneath the cover without precipitating a break-up and, after subsequent additional ablation of the cover, be followed by a thermal break-up with minimal backwater. The greatest impact of an elevated ice cover is likely to be on middle order events: ones that regularly used to flood the delta. For such events, it would seem reasonable to assume that the effectiveness or potency of tributary flow would be diminished because of an elevated cover. Although the ultimate severity of a particular event would depend on a number of the other controlling factors described above, over the longer term, some reduction in the recurrence interval of this order of break-up event should occur. This also offers a partial explanation of why there has been a decrease in the frequency of flood level break-up events after regulation.

**SUMMARY**

Previous work by Prowse and Lalonde (1996) has established that river ice break-up has been responsible for flooding of the perched basin environment of the PAD and that such events noticeably diminished in the mid-1970s. This study evaluated a suite of hydrometeorological conditions that control the severity of break-up and might explain the reasons for the significant temporal change in the frequency of major flood events. Special attention was placed on the effects of flow regulation and climate variability.

In the lower portions of the Peace River, flow regulation seems to have produced minor changes in factors, such as ice thickness and strength, that could significantly affect the severity of break-up and related ice-jam flooding. Temporal analysis of these factors, however, also detected a weak climate signal suggesting that since approximately the mid-1970s the period of ice cover may have become slightly warmer and the pre-break-up melt period may have become more intense and/or more protracted. Although this needs to be explored more thoroughly, such factors could also favour the development of thermal over dynamic break-ups, and hence reduce the probability of severe ice-induced flooding.

A common perception was that reduced flows resulting from regulation were responsible for the decline in break-up severity and related flooding. Results show, however, that flow contributed from above the dam is
higher on average at the time of break-up near the Peace–Athabasca Delta in the post-regulation period than it was prior to regulation. The major ice break-ups that occurred in the 1960s prior to regulation and in the early 1970s after regulation have been associated with large runoff events from downstream tributaries, especially the Smoky River. The flow contributed by tributaries at the time of break-up far exceeds that contributed by headwaters above the point of regulation. These large tributary flow events also appear to be correlated with large spring snowpacks and associated snowmelt runoff. A preliminary evaluation of temporal trends in the size of the snowpack on the Smoky River suggests that there has been a shift, in the mid-1970s to values lower than the long-term average. A similar trend has been identified in British Columbia and appears to be responsible for decreased spring runoff on some rivers.

The major effect of regulation on the severity of break-up near the Peace–Athabasca Delta is related to the higher winter flows and freeze-up elevations. In general, the higher a freeze-up cover is stabilized, the greater the flows it can pass without breaking. The amount that spring flows exceed a freeze-up level depends on two contributing sources: the upstream flow from above the point of regulation and the downstream tributary flow. Under regulated conditions, a major increase in upstream flows (above the point of regulation) is unlikely at the time of break-up near the Peace River Delta under the standard operational strategy of the upstream reservoir, i.e. at the time of transition to lower summer releases. Furthermore, if the amount of regulated flow at the time of break-up is also declining, additions from tributary flow will also have to account for this ‘loss’ to the main stem discharge. Thus, under the current regulated regime, production of severe break-ups has become more dependent on tributary inflow, particularly from the Smoky River. Large spring runoff from the tributaries has been effective, since regulation, in producing large break-up floods (e.g. 1972 and 1974) but the apparent decline in spring snowpacks has reduced their subsequent effectiveness.

Future research should focus on climatological explanations for the changes in the magnitude of winter snowpacks and on obtaining site specific data near the PAD for the modelling of ice-jam events.

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REFERENCES


APPENDIX A. MODELLING OF SPRING ICE ABLATION

Energy balance model

The energy balance approach was employed by applying the law of conservation of energy to a control volume of ice (e.g. Gray and Prowse, 1993). The upper and lower boundaries of the control volume were the ice–air and ice–water interfaces, respectively. Horizontal transfers of energy were assumed to be negligible. The total energy available for melt ($Q_m$) is expressed as

$$Q_m = S \downarrow (1 - \alpha_i) + \left(L \downarrow + L \uparrow\right) + Q_h + Q_e + Q_p + Q_w - \Delta U/\Delta t$$  (A1)

where $S$ = global (short-wave) radiation; $\alpha$ = albedo of the surface ice layer; $L$ = atmospheric long-wave radiation; $L_s$ = surface long-wave radiation (in subsequent discussion net short-wave and net long-wave radiation are noted as $S^*$ and $L^*$, respectively, and $Q^*$ as their net sum); $Q_h$ = sensible energy, the turbulent flux of energy exchanged at the ice surface due to a difference in temperature between the ice surface and overlying air; $Q_e$ = latent energy, the turbulent flux of energy exchanged at the ice surface owing to vapour movement as a result of a vapour pressure difference between the ice surface and the overlying air (evaporation represents a loss and condensation a gain); $Q_p$ = advected energy, energy derived from external sources, such as precipitation, that is added to the volume; $Q_w$ = hydrothermal energy, the turbulent flux of energy exchanged at the ice bottom owing to a difference in temperature between the ice bottom and the underlying water (this term is comprised of heat owing to fluid friction, geothermal heat, heat from groundwater flow and bed sediment heat); $\Delta U/\Delta t$ = rate of change of internal energy in the volume per unit surface area per unit time; all terms except $\alpha$ which ranges from 0 to 1.0, are in W/m$^2$.

The sensible and latent heat fluxes were modelled using bulk aerodynamic formulae that have been proven in many snow and ice ablation studies to provide good daily results (e.g. Moore, 1983). Following the approach of Price and Dunne (1976), a stability factor was used for non-neutral conditions (stable and unstable). Since, as outlined below, calculations were performed only for the major pre-break-up melt period, the storage term was treated as zero. No calculation of the hydrothermal flux was made because of the lack of appropriate data and because it was assumed to be small relative to the atmospheric fluxes during the pre-break-up period (e.g. Prowse and Marsh, 1989). Herein, the total atmospheric fluxes are symbolized by $Q_a$ (i.e. $Q_m + Q_w$). In calculating $Q_p$, wet-bulb air temperatures were extracted from temperature and humidity records and used as a surrogate of the precipitation temperature (e.g. Gray and Prowse, 1993). All calculations were performed at an hourly time step and summed over the respective melt periods.

Degree–day model

For years when there were insufficient data to adopt an energy balance approach, a simplified temperature-index approach was employed (e.g. Bilello, 1980). Reductions in ice thickness were estimated according to

$$\Delta h_i = \tau T_{\Sigma M}$$

(A2)

where $\tau$ = an empirical coefficient (m/$^\circ$/C/d); $T_{\Sigma M}$ = accumulated thawing degree–days (°C/d) with a base of $-5^\circ$C. Bilello (1980) found $\tau$ to vary from approximately 0-004 to 0-01 for northern (i.e. $> 60^\circ$N) Canadian and Alaskan rivers, and Prowse et al. (1989) found the lower value to be suitable for a temperate river in eastern Canada.

Data/model requirements for calculations of ice ablation

The closest meteorological station to the PAD is at Fort Chipewyan where recordings of temperature, wind speed, relative humidity, barometric pressure, snow cover and some sky conditions (e.g. cloud height, type and opacity) are made. The next closest station is located at Fort Smith, NWT (Figure 1), approximately 150 km to the north-west. Hourly data are available for Fort Chipewyan for the period 1963–1979 but only daily data since 1979. These data were sufficient for the degree–day model and to calculate the convective fluxes in the energy balance model, but no direct measurements were available for solar radiation. Given the importance of short-wave radiation to melt and ice decay (e.g. Ashton, 1985; Prowse et al., 1990a,b), a cloud-layer model (‘McMaster Model’) was used to derive $S$ (Davies and McKay, 1982; Davies et al., 1984). Since the study area was considered not to experience significant effects of aerosol attenuation (i.e. non-urban), an aerosol transmission value of 1-0 was employed. A single scattering albedo value of 0-75 was also used (J. A. Davies, personal communication). Approximately 22 meteorological variables were required as input (see Prowse et al., 1996a, for details). Testing of the model was conducted using data recorded at nearby Fort Smith over the period 1972–1978 and produced an average correlation coefficient of 0-91. Experience operating the model indicated that it was most appropriate to use only the first layer of
cloud data, as it provided the most consistent results. A value of 0.5 for $x_5$ was considered to best represent the melting snow and ice-covered conditions that would have dominated during the calibration period of April–May. Following final validation, the modified McMaster Model was used to predict hourly global radiation data using Fort Chipewyan data for the spring break-up periods in the years 1963 to 1979.

The absence of long-wave radiation data for Fort Chipewyan also meant that $L_\downarrow$ and $L_\uparrow$ had to be modelled. $L_\uparrow$ was determined following the method outlined in Brutsaert (1982) with an emissivity of the melting surface of 0.98. Since ice temperatures were unavailable, the river ice surface was assumed to be at 0 °C if the air temperature was above 0 °C. Otherwise the surface temperature was assumed to be equal to the near-surface air temperature. The atmospheric long-wave radiation was calculated similarly, but atmospheric emissivity was determined from a method outlined by Marks (1979) and corrected for barometric pressure and temperature differences with elevation. Values for clear sky $L_\downarrow$ were adjusted to account for the effect of varying cloud type and coverage (Sellers, 1965).

**Period of ice ablation**

Unless there is a large hydrothermal flux, ice ablation does not usually commence until melt of the surface snow cover has begun. At this point, melt may occur at the ice surface owing to atmospheric exchanges and at the base because of terrestrial runoff, groundwater and radiative heating. Calculation of the various heat fluxes required knowledge of the cover during its melt state, so that proper assumptions could be made about surface conditions, such as temperature and vapour pressure. A procedure was developed, therefore, to identify the period of active melt, i.e. the melt period spanning the initiation of pronounced spring ice melt to the day of break-up. The length of this interval and the accumulated heat flux that occurs within it are indicative of the intensity of the pre-break-up melt period, as discussed earlier regarding types of break-up. Three criteria were employed in combination to define the date of ablation initiation. These included daily snow depth from land-based stations, daily maximum and minimum air temperature and small stream runoff.

Daily snow depth records were assembled from both the Fort Chipewyan and Fort Smith climate stations. Daily maximum and minimum air temperatures were also assembled from the Fort Chipewyan climate station. Daily stream flow records were assembled for seven nearby, small (<10 000 km²) streams. It was reasoned that when these basins began to produce spring flow, the ablation period had begun and the ice cover would also be in a state of ablation. It was further assumed that the relatively shallow snowpack on the wind-swept ice cover would ablate in the early stages of melt recorded at the land-based sites and that the above defined ablation period would be characterized primarily by ice ablation at river sites.

**APPENDIX B. MODELLING OF CHANGES TO ICE STRENGTH**

Analytical procedures first involved modelling of incoming short-wave radiation for each period of pre-break-up melt. The limitations of available meteorological data restricted the analysis to the years 1963–1979. In calculating net short-wave radiation entering the cover, a surface albedo of 0.4 was used, considering that this best approximates melt conditions typical of both granulating white ice (frazil accumulations) or candling black ice (thermal ice) (Prowse and Marsh, 1989). Radiation attenuation within the ice sheet was modelled after the method outlined by Ashton (1985) and modified to account for spectral selectivity with depth using a power law dependence for first-year blue (black) ice established by Grenfell and Maykut (1977). An average ice sheet thickness of 0.95 m was employed. The melt fraction or porosity values were then calculated and the porosity values used for quantifying a strength index according to

$$
\theta/\theta_0 = 1 - c\Phi^k
$$

where $\theta/\theta_0$ is the ratio of strength at any porosity development to original strength value at 0 °C. The constants $c$ and $k$ were assigned values of 2.8 and 0.5, respectively, which were believed to describe best the typical grain geometry and range of melt conditions (e.g. see Prowse and Demuth, 1992).