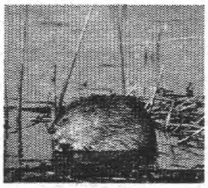
Canada Alberta Northwest Territories

Northern River Basins Study





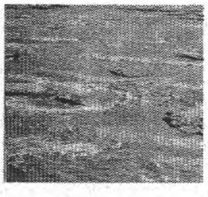





NORTHERN RIVER BASINS STUDY PROJECT REPORT NO. 122

THE EFFECTS OF FLOW REGULATION ON FREEZE-UP REGIME

PEACE RIVER, TAYLOR TO THE SLAVE RIVER






Prepared for the
Northern River Basins Study
under Project 1423-C1

by

D. D. Andres
Trillium Engineering and Hydrographics Inc.

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PREFACE:

The Northern River Basins Study was initiated through the "Canada-Alberta-Northwest Territories Agreement Respecting the Peace-Athabasca-Slave River Basin Study, Phase II - Technical Studies" which was signed September 27, 1991. The purpose of the Study is to understand and characterize the cumulative effects of development on the water and aquatic environment of the Study Area by coordinating with existing programs and undertaking appropriate new technical studies.

This publication reports the method and findings of particular work conducted as part of the Northern River Basins Study. As such, the work was governed by a specific terms of reference and is expected to contribute information about the Study Area within the context of the overall study as described by the Study Final Report. This report has been reviewed by the Study Science Advisory Committee in regards to scientific content and has been approved by the Study Board of Directors for public release.

It is explicit in the objectives of the Study to report the results of technical work regularly to the public. This objective is served by distributing project reports to an extensive network of libraries, agencies, organizations and interested individuals and by granting universal permission to reproduce the material.

This report contains referenced data obtained from sources external to the Northern River Basins Study. Individuals interested in using external data must obtain permission to do so from the donor agency.

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Whereas the above publication is the result of a project conducted under the Northern River Basins Study and the terms of reference for that project are deemed to be fulfilled,
IT IS THEREFORE REQUESTED BY THE STUDY OFFICE THAT;
this publication be subjected to proper and responsible review and be considered for release to the public.

F. J. Wrona — 4 Apr 96
(Dr. Fred J. Wrona, Science Director) — (Date)

Whereas it is an explicit term of reference of the Science Advisory Committee "to review, for scientific content, material for publication by the Board",
IT IS HERE ADVISED BY THE SCIENCE ADVISORY COMMITTEE THAT;
this publication has been reviewed for scientific content and that the scientific practices represented in the report are acceptable given the specific purposes of the project and subject to the field conditions encountered.
SUPPLEMENTAL COMMENTARY HAS BEEN ADDED TO THIS PUBLICATION: [ ] Yes [ ] No

P. A. Larkin — 15 April 96
(Dr. P. A. Larkin, Ph.D., Chair) — (Date)

Whereas the Study Board is satisfied that this publication has been reviewed for scientific content and for immediate health implications,
IT IS HERE APPROVED BY THE BOARD OF DIRECTORS THAT;
this publication be released to the public, and that this publication be designated for: **[ ] STANDARD AVAILABILITY [ ] EXPANDED AVAILABILITY**

L. Partington — 21 April 1996
(Lucille Partington, Co-chair) — (Date)

R. McLeod — 10 April 1996
(Robert McLeod, Co-chair) — (Date)

THE EFFECTS OF FLOW REGULATION ON FREEZE-UP REGIME PEACE RIVER, TAYLOR TO THE SLAVE RIVER

STUDY PERSPECTIVE

One of the questions posed by the Northern River Basins Study Board requested scientists to assess the effects of flow regulation on the aquatic/riparian ecosystem. Attention was focused on the Peace River system because of the scale of flow regulation (Bennett dam) and its effects on the Peace mainstem and areas downstream. While the Peace - Athabasca Delta has been studied since the early 1970's, the effects on the Peace River, especially on ice processes, have received limited study.

> ***Related Study Questions***
>
> *10. How does and how could river flow regulation impact the aquatic ecosystem?*

This report describes the effects of flow regulation on the freeze-up processes of the Peace River. Current records indicate that flow regulation has increased the winter discharge by a factor of two or three, causing a significant impact on the ice characteristics of the river. This project focused on using ice modelling and reviewing ice observations and other field data to characterize the extent of change and its effects on ice in the river channel. The effects of flow regulation on the ice regime were determined to include:

1. higher water elevations (up to six metres), throughout the winter in locations where a consolidated ice cover forms (mainly upstream of Manning);
2. a reduction in the duration of the ice cover upstream of Fort Vermilion (down to zero immediately downstream of the dam);
3. losses of up to 30% of the flow into storage as the ice cover advances (may affect calculations of minimum flows);
4. increased frazil ice production due to the open water downstream of the Bennett dam all winter; and
5. thicker deposits of frazil in low velocity areas upstream of the Vermilion chutes (may reduce or eliminate flow in some shallow areas habitat).

The results of this study will be compiled with material from other NRBS investigations (Proceedings of the IFN Workshop, Channel Morphology and Riparian Vegetation on the Peace River within Alberta, Regulation Effects on the Slave River Delta: Landform and Distributary Sensitivities to Changes in River Regime and Aquatic Habitat Mapping for Instream Flow Needs Analysis - Peace River Pilot Project) and outside sources to provide a comprehensive assessment of the effect of flow regulation on the aquatic/riparian ecosystem.

REPORT SUMMARY

This report has reviewed the processes by which an ice cover forms on large regulated and non-regulated rivers. Explicit equations and algorithms have been presented that quantify these processes. Work that had been undertaken previously on the Peace River was also described to provide a framework for the calibration of these algorithms for the Peace River in both its regulated and non-regulated condition. The significant theoretical advances that were made include the development of a procedure to forecast freeze-up on a non-regulated river and the derivation of a stability relationship that uses both air temperature and discharge to determine whether a juxtaposed or consolidated ice cover will form. The latter development is important to characterize the type of ice cover that will occur on the Peace River under regulated conditions.

In addition, the hydraulic characteristics of the Peace River were evaluated for six distinct reaches between the Slave River and Taylor using the existing data base. The climatological characteristics of the basin were summarized, along with a description of the spatial and temporal variation in the flows for the periods before and after regulation.

Prior to regulation, at flows of less than 1000 m^3/s, the river cooled from a maximum annual water temperature of about 22°C to 0°C at the same rate as the declining air temperature Ice began to form in early November in most years, and an ice cover formed by multiple lodgements when the surface ice concentration neared 100% and the discharge decreased sufficiently to reduce the width of the flow by about 10%. A stable ice cover usually formed in early November at Peace Point and in late November or early December at Peace River. There is no data for Taylor, although the freeze-up probably occurred in early December. The ice thickness associated with this type of freeze-up generally ranged from 0.5 to 1.0 m. The stage increase was typically between 1.0 to 2.0 m. In some cases, due to declining flows during and following the formation of the ice cover, the stage decreased after the ice cover was established.

Since regulation, the discharges are, on the average, about two to three times greater than those prior to regulation. This high discharge of relatively warm water from upstream of the has delayed the time of freeze-up and shortened ice duration of the ice cover significantly in the reaches upstream of Fort Vermilion. At Taylor, and upstream of the BC/Alberta border, an ice cover is an exception rather than a rule. At Peace River, and downstream to Fort Vermilion, the freeze-up date has been delayed by as much as one to two months. Only minor effects due to regulation are evident on the freeze-up ice regime downstream of the Vermilion Chutes and at Peace Point.

After regulation, the ice cover downstream the Notikewin River generally forms by juxtaposition due to the very mild slopes. The ice cover thickness in these two reaches is only about 0.5 m thick, immediately after freeze-up and the stage increase associated with freeze-up is only about 1 to 2 m. The increase in stage is due mostly to the additional flow resistance of the ice cover. In the reaches between the Notikewin River and Dunvegan, where higher slopes are evident, either a juxtaposed or consolidated ice cover can form. For typical post-regulation discharges, the air temperature must be at least -30°C for a juxtaposed cover to form. To ensure that a juxtaposed ice cover forms, regardless of the air temperatures expected, the discharge should be less than 800 to 1000 m^3/s. The stage increase under a juxtaposed ice cover is less than 2 m, but for a consolidated ice cover the stage increase can be as great as 5 m, with an ice thickness of about 4 m. Between Hudson Hope and Dunvegan, the steeper river slopes prevent the formation of a juxtaposed ice cover for any reasonable combination of discharge and air temperature. Although the development of an ice cover in these two reaches is infrequent and when it does occur its duration is short lived, the formation thickness can approach 5 m and the increase in the stage can be up to 6 m.

The main physical impacts on the environment relate primarily to (1) the existence of high water levels for long periods of time in areas where a consolidated ice cover has developed, (2) the losses in up to 30% of the flow into channel storage as the ice cover advances, (3) the potential unstable water levels and ice thicknesses that are evident within 100 km of the advancing ice cover, (4) the reduction in the duration of an ice cover for most of the length of the Peace River, and (5) dramatically thicker deposits of frazil in low velocity areas of the river upstream of the Vermilion Chutes.

Although algorithms have been developed for many of the process identified on the Peace River, additional work is required to improve the modelling capabilities. Additional observations and measurements need to be carried out downstream of the Vermilion Chutes to characterize better the freeze-up process in that reach. It is also suggested that bench marks established around the Vermilion Chutes as part of this study be referenced to a common datum. This will improve the understanding of the hydraulics of the Chutes. From a modelling point of view, more work is required to verify the stability criteria used in determining the dominant mode of cover formation. An important component of this work will be the unsteady simulation of a consolidation event. Also, some effort must be expended to explicitly model the formation of frazil ice floes.

ACKNOWLEDGMENTS

This work described in this report was jointly funded by the Alberta Research Council and the Northern River Basins Study. The author acknowledges the assistance of T. Prowse and J. Choles in the development of the study plan. Grateful acknowledgment is made of data provided by Environment Canada. V. Elder has been particularily helpful throughout the course of the work. G. Fonstad of Alberta Environmental Protection also provided important data. Finally, Bernard Trevor of Trillium Engineering and Hydrographics Inc. assembled most of the hydraulic and ice related data.

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1.0 INTRODUCTION

1.1 BACKGROUND

Since the construction of the Bennett Dam, the hydrologic regime of the Peace River has been changed dramatically. Much higher flows at freeze-up have altered the timing and characteristics of the formation of the ice cover. Generally, the onset of freeze-up has been delayed and a much thicker and more variable ice cover forms over much of the river downstream of Taylor. During breakup, the effects of the Dam tend to moderate the breakup process, largely due to the creation of a heat sink immediately downstream of the dam and a reduction in the amount of ice that can contribute to the development of ice jams downstream of Peace River town. The result is that, in many of the reaches, high ice-related spring water levels are no longer produced and/or the annual peak water levels now occur during the freeze-up period instead of the breakup period.

A number of changes to the regime of the Peace River, and more importantly to the Peace/Athabasca Delta, have been observed over the last number of years. The Delta appears to be drying out. Furthermore, these changes seem to have a direct impact on the biological productivity of the Delta and this is causing concerns. There is substantial confusion over why these changes have occured. The effects of regulation and the changes in the climate regime of the recent winters have been cited as being the possible causes for the hydrologic trends in the Delta. There is a need to resolve these issues within the context of an improved understanding of the ice related processes on the Peace River. Unfortunately, up to now, only limited work has been done to quantify the freeze-up and breakup processes so that these issues can be addressed objectively.

Currently, a number of agencies are examining the hydraulic characteristics of the Peace River, under both open water and ice covered conditions. These agencies include Alberta Environmental Protection, BC Hydro, the Northern Rivers Basin Study Board, and the Alberta Research Council, under contract to Environment Canada. Unfortunately, much of this work is addressed towards specific issues and no effort has been made to combine various ice observations with the hydrometeorological records, and the hydraulic characteristics of the river to identify the nature of the freeze-up processes and to understand the implications of the freeze-up processes on the ultimate hydrological and biological regimes of the Peace River and the Peace/Athabasca Delta. This project described herein will undertake some of this work.

1.2 STUDY OBJECTIVES

The objectives of this work are to establish a framework to quantify the importance of the freeze-up process on the hydrologic and biologic regimes of the Peace River. The work described herein identifies the relevant processes which produce a stable ice cover on the river, compares the pre-dam and post-dam freeze-up regime on a reach by reach basis, identifies the impact of regulation on the timing of freeze-up over reaches of interest, and determines the boundary and initial conditions for the subsequent (or concurrent) analysis of the breakup processes.

The following sections of the report work will include:

1. A description of the study area with respect to its location, physiography, and political boundaries;

2. A summary of the mechanics of freeze-up, along with a dimensionless relationship that identifies the dominant ice cover characteristics that would occur under regulated conditions for a particular combination of air temperatures and discharges within a channel with particular hydraulic characteristics;

3. A summary of the hydraulic characteristics of the relevant identifyable reaches located between the Delta and Taylor;

4. A summary of the pre-regulation and post-regulation flow characteristics for the freeze-up period for each year for which records exist and for locations where sufficient hydraulic information is available; and

5. A summary of the freeze-up characteristics (discharge, stage, air temperature, type of ice cover, etc.) at each of the salient hydrometric gauges;

From the above, the dominant freeze-up modes (as a function of discharge and air temperature) will be identified for each of the reaches. The effects of regulation on the ultimate character of the ice cover will then be assessed. Finally, attempts will be made to extend the freeze-up characteristics to address issues related to ground water levels and habitat adjacent to the river.

2.0 STUDY AREA

2.1 LOCATION

The study area is the Peace River between the Taylor and the Slave River (Figure 1) in northern British Columbia and Alberta. The flow in the river is controlled by releases from the Bennett Dam. The river drains the eastern slopes of the Rocky Mountains, and flows eastward from Hudson Hope, B.C., crossing the Alberta-British Columbia border at Clayhurst. At Peace River town, the course of the river changes to a northerly direction, finally entering the Slave River downstream of Peace Point (Figure 1). The Peace River has been gauged at a variety of locations since 1915, and more intensely following the construction of the Bennett Dam. In Alberta, the most relevant active gauging stations, which provide data to analyse the ice processes for this study, are those at Taylor (#07FD002), Dunvegan (#07FD003), Peace River (#07HA001), and Peace Point (#07KC001). In addition to those stations, inactive stations located at Fort Vermilion (#07Hf001) and at Carcajou (#07HD001) will be used to quantify the channel characteristics. It should be noted that the gauges at Dunvegan, Carcajou, and Fort Vermilion operate or have operated only in the summer months, and thus are not used to characterize the ice regime.

2.2 CLIMATE

As shown in Figure 1, the study reach spans an extremely large area. Fortunately this area is relatively homogeneous with respect to the meteorological conditions on any one given day. There are some variations with both latitude and longitude, with the latitudinal variations being more significant. Table 1 summarizes the ranges of the monthly means and extremes for the entire basin.

Table 1 Summary of monthly mean and extreme air temperatures in the basin, 1961-1990

Month	Temperature[1] (°C)		
	Mean	Minimum	Maximum
August	14.8	8.6	20.9
September	9.2	3.5	14.9
October	3.3	-1.6	8.1
November	-8.7	-12.9	-4.6
December	-16.0	-20.6	-11.6
January	-18.4	-23.2	-13.7

[1] Average for Fort St. John, Peace River, and Fort Vermilion.

In November, when freeze-up typically starts, the mean monthly air temperature is in the order of -10°C. At the peak of the winter the mean monthly air temperature varies from -20 to -25°C. The climatic normals (1961 to 1990) for the basin suggest that there are on the average about 100 to 140 days in the year when the mmaximum daily air temperature is less than 0°C. Typically, there is about 1700 to 2500°C-days of accumulated freezing in the basin for any given year. Figure 2 illustrates the variability from year to year over a particularily warm period between 1977 and 1994.

For simplicity, three meteorologic stations - Fort St. John, Peace River, and Fort Vermilion can be used to quantify the daily air temperature and solar radiation characteristics for the purposes of modelling the ice generation and the formation of an ice cover under regulated conditions. It should be recognized that the Fort St. John and Peace River are more relevant in determining ice production in upper reaches of the Peace River; and the Fort Vermilion data are more relevant for computing the rate of ice growth in the juxtaposed ice cover downstream of Manning.

2.3 DISCHARGE REGIME

The gauge at Taylor defines the inflow from the upper basin into Alberta, while the gauge at Peace Point defines the outflow from entire basin into the Northwest Territories. During the winter the main control of flow to the system is the Bennett Dam. Most of the inflow downstream of the Dam originates in the Pine, Smoky, and Wabasca Rivers.

2.3.1 Spatial Variability

Table 2 summarizes the mean monthly flows on the Peace River and its tributaries for the pre- and post-regulated periods. Obviously, the tributaries are not regulated so there will be no change in their flow patterns following regulation of the Peace River. Prior to regulation the mean monthly flow along the Peace River varied from about 1200 m^3/s in the month of October to about 1540 m^3/s at Peace Point; an increase of about 340 m^3/s. Most of this inflow originated from the four largest tributaries: the Beaton, Kiskatinaw, Smoky, and Wabasca Rivers. By January, the flow in the Peace River at Taylor had dropped to about 330 m^3/s and the flow at Peace Point similarily decreased to about 500 m^3/s. For that month the four main tributaries made up only about 40% of the difference. In January, the inflow between Taylor and Peace Point is composed proportionately more from the distributed lateral inflow from the various small basins than from the four main sources of inflow.

2.3.2 Effects of Regulation

The river has been regulated since 1972 by the Bennett Dam. This has substantially increased the discharge during the winter period. For example at Peace River, the annual mean flow is about 1800 m^3/s and under natural conditions the winter flows were in the order of 200 to 500 m^3/s. However since regulation, the winter flows have been substantially increased into the range of 1000 to 2000 m^3/s (Figure 3). Also, the natural flow recession in the river during the winter period has been eliminated and the flow is at a relatively high discharge over the entire winter period. Under regulation, the proportion of the flow in the river that originates from the tributaries has decreased from about 20 to 30% to between only 10 and 20%. In fact, the hydraulic characteristics after regulation are dominated by the inflows from the Bennett Dam, so that for all practical purposes the local inflows can be neglected, for the most part, in any freeze-up modelling.

Table 2 Comparison of pre- and post-regulation freeze-up discharges on the Peace River

River	Mean monthly discharge (m^3/s)[1]			
	October	November	December	January
Peace River at Taylor	1200 (1400)[2]	790 (1450)	395 (1420)	330 (1370)
Beaton River	18.9	7.23	2.46	1.10
Kiskatinaw River	4.77	2.74	1.55	0.936
Smoky River	233	114	63.1	52.7
Peace River at Peace River	1440 (1630)	932 (1540)	467 (1430)	390 (1430)
Notikewan River	4.65	1.68	0.449	0.183
Wabasca River	76.5	41.1	22.8	15.8
Peace River at Peace Point	1540 (1831)	1140 (1600)	540 (1530)	501 (1770)

[1] Discharges based on period of record between 1958 and 1990.
[2] The bracketed values indicate the post-regulated period.

3.0 CHANNEL CHARACTERISTICS

The hydraulic information available for the Peace River is dispersed in the literature. Data from Kellerhalls, Neill, and Bray (1972) were used to characterize the various sub-reaches between Hudson Hope and the Slave River. This data was augmented from a variety of other sources, including Alberta Research Council (ARC) records, Alberta Environmental Protection (AEP) records, and from recent work carried out by for the NRBS by ARC and Northwest Hydraulic Consultants Ltd. A slope profile from the various water level data is shown in Figure 4. The extent of six reaches with more or less homogeneous hydraulic characteristics was chosen according to the average slope and the sinuosity. The average slope for each reach was determined from both average map slopes and locally surveyed slopes.

The channel geometry in each sub-reach is described by a generalized cross section based on a composite of all the available cross sections in that reach. The typical adopted channel cross section for each of the individual reaches is shown in Figure 5. The relationship between discharge, top width, and mean depth was calculated from the reach-average slope and the Manning's roughness evaluated from site-specific measurements. Table 3 summarizes the data illustrated in Figures 6 to 10.

Table 3 Summary of adopted hydraulic characteristics of individual reaches

Reach	1	2	3	4	5	6
Location	Slave River to Vermilion Chutes	Vermilion Chutes to Notikewin River	Notikewin River to Daishowa	Daishowa to Dunvegan	Dunvegan to Taylor	Taylor to Hudson Hope
Distance from mouth (km)	0 to 330	330 to 640	640 to 798	798 to 970	970 to 1123	1123 to 1219
Sinuosity	1.23	1.48	1.48	1.12	1.10	1.10
Meander length (km)	17	14	10	11	16	-
Slope	0.000075	0.000065	0.00025	0.00028	0.00037	0.00049
Manning's roughness[1]	0.031	0.034	0.045	0.045	0.030	0.041
Channel geometry[1]						
Top width (m)	590	460	410	410	420	420
Mean depth (m)	2.78	3.53	2.64	2.64	2.28	2.42
Mean velocity (m/s)	0.55	0.55	0.67	0.67	1.13	0.98

[1] The hydraulic characteristics are tabulated for a discharge of 1000 m^3/s.

4.0 FREEZE-UP PROCESSES

The nature of freeze-up processes on rivers depends primarily upon meteorologic conditions and hydraulic characteristics of the river. Most large rivers can be considered as steep rivers, i.e., rivers in which the velocity is large enough to prevent a surface cover from forming by other than through an accumulation of frazil. That is, frazil production is the major ice production mode. Skim ice, when it forms either contributes to the formation of border ice or it is entrained into the flow and mixed with the frazil.

To model the freeze-up processes, it is necessary to quantify or mathematically represent the heat loss from both the water and the ice surfaces, the production of frazil, the lodgement mechanisms, the border ice growth and the development of a stable ice cover. Some of these processes are better known than others. For example, more is known about heat loss, frazil production, and cover formation than is known about lodgement and border ice growth.

This section will describe the various segments of the freeze-up process and document or summarize some of the relevant equations which can be used to quantify the production of ice and the development of a stable ice cover.

4.1 HEAT LOSS

The thermal regime of a natural stream is dependent upon many factors, principal among them being the stream geometry, the meteorological regime of the area, and the quantity of heat introduced by tributary inflows. For a natural stream without imposed, artificial heat loads, the various meteorological factors may combine to produce either a net influx of heat, resulting in an increase in water temperature or a net afflux of heat and a decrease in water temperature. The former generally occurs during warm summer periods or during the day, while the latter is characteristic of cold weather conditions or at night.

The representation of stream temperatures and the production of frazil requires an evaluation of the heat transfer (flux) across the stream boundaries. These boundaries include both the stream surface (air-water interface) and the stream bed. Heat transfer at the water surface is extremely large in comparison to heat transfer through the stream bed. Four main mechanisms can be identified as contributing to the total heat flux, $\mathbf{H}_t$. They are short wave solar radiation, longwave radiation, evaporation/condensation, and convection. Other components such as melting snow, precipitation, and geothermal energy can also be considered. However, these are generally much smaller than the preceding components or occur sporadically in time. They are not considered in the analysis contained herein. The total heat flux can be divided into two main components: the solar radiation component, $\mathbf{H}_s$ and the temperature related heat transfer component, $\mathbf{H}_a$ which is the sum of the longwave radiation, evaporation/condensation, and convection.

Direct measurements of the solar (shortwave) radiation at the site under consideration are obviously the best method for determining the magnitude of the energy supplied by this means. In cases where these measurements do not exist, measurements from stations at some distance from the site can be substituted, provided that there is reason to believe that the cloud cover conditions at the two sites are comparable. If direct measurements of shortwave radiation are not available, estimates of its magnitude can be made by using tables, graphs, or formulas relating percentage of possible sunshine or cloud cover to the percentage of possible solar radiation that reaches the ground. Bolsenga (1964) and Gray (1970) have provided tables of daily and monthly sums of clear sky radiation for various latitudes and air masses. The effects of cloud cover can be accounted for by a relationship initially suggested by Angstrom (List, 1963)

[1] $$H = H_{cs}[0.35 + 0.061(10 - C)]$$

where H_{cs} is the incoming clear sky radiation, **C** is the cloud cover expressed in tenths, and **H** is the incoming shortwave or solar radiation at the surface of the earth. This parameter can also be computed from the more commonly measured daytime hours of bright sunshine, **h** if a correlations between the two can be obtained. For northern Alberta, Andres (1988) suggested that

[2] $$\frac{H}{H_d} = 0.65\left(\frac{h}{h_{sm}}\right) + 0.39$$

where H_d is the maximum daily clear sky solar radiation and h_{sm} is the daily maximum hours of bright sunshine. Both H_d and h_{sm} are a function of the latitude and declination of the sun and hence can be computed explicitly for any day at any location. Other more sophisticated methods are available to calculate the solar radiation, however their use requires substantially more meteorologic data. The additional data does not necessarily improve the accuracy of the estimate and hardly justifies the effort.

The net incident solar radiation is also a function of the aspect of the water surface and of any obstruction around the stream surface. Some rather tedious (but not complex) methodologies have been suggested to compute the effects of vegetation or shading from valley walls on the incident solar radiation. These require considerable knowledge of the heights of the valley walls and the aspect of the water body. This procedure is not often used; instead, the reflected shortwave radiation can computed by calculating or calibrating an exposure index which takes into account all potential losses (Andres, 1988). Thus,

[3] $$H_i = \Gamma H$$

and

[4] $$H_s = (1 - \alpha_w)\Gamma H$$

where α_w is the albedo, H_i is the incident shortwave radiation, H_s is the net shortwave radiation absorbed at the surface of the water body, and Γ is the exposure factor used to quantify the effects of the shading from valley walls.

It is convenient to combine all the temperature-related heat flux terms and linearize them in the form of Equation [5],

$$H_a = h_w (T_a - T) \qquad [5]$$

where T_a and T are air and water temperatures, respectively, and h_w is a heat transfer coefficient which accounts for wind, barometric pressure, relative humidity, and cloud conditions. This linear approximation is acceptable as a first approximation because many of the above noted effects cannot be accounted for, even for the most rigorous analysis.

The value of h_w for a particular location can be determined by calibration and is often used to describe all the temperature-related heat transfer processes by one lumped coefficient. Unfortunately, many investigators, Ashton (1983) for example, also include the effects of the solar radiation in their definition of the convective heat transfer coefficient. Thus, they probably overestimate its true value. Andres (1984), for the Upper Peace River, found that after separating out the effects of solar radiation, the convective heat transfer coefficient above water, h_w was in the order of 15 W/m^2-°C.

The exposure factor, Γ is not a constant, but varies depending upon the angle of the sun and configuration of the stream in relation to the characteristics of the valley walls. Andres (1988) suggested that during March and April, the exposure factor was about 0.75 for the Athabasca River between Athabasca and Fort McMurray.

The surface albedo is a function of the solar angle and characteristics of the surface. Anderson (1954) summarizes some investigations which evaluated the albedo of water. He suggests that the solar radiation is reflected from both the water surface and a stratum of relatively opaque water composed of bubbles and suspended material located just below the surface. The reflectivity also tends to increase when the surface is hydrodynamically rough and under cloudy conditions. However, it is never more than 0.09 for solar angles greater than 30°, and under overcast skies it averages about 0.08. Anderson found that for a sun angle of 30°, which is greater than that in Alberta during the freeze-up period, the albedo varied between 0.05 and 0.12, but typically has values of about 0.075 for a variety of cloud conditions, from scattered to broken to completely overcast.

Adding Equations [4] and [5] gives the net heat flux at the air water interface, which is given by

$$H_t = (1 - \alpha_w)\Gamma H + h_w (T_a - T) \qquad [6]$$

4.2 FREEZE-UP ON A REGULATED RIVER

The production of ice in a flowing stream occurs in two modes, border ice growth and frazil production. In most cases border ice is first to appear. It forms in low velocity areas adjacent to the river bank where vertical mixing is insufficient to entrain the ice which forms on the surface of the flow. Frazil production begins once the entire column of water cools to a supercooled condition. This requires considerably more heat loss than thermally stratified flow and as a result frazil generally appears well after border ice has formed.

4.2.1 Border Ice Growth

The growth of border ice occurs in two modes. Initially a lateral growth occurs due to a lateral heat loss through the ice and into the bank. This is accompanied by a thermally induced thickening of the ice cover. The result is a smooth ice cover extending out into the flow. This ice cover has the greatest thickness at the bank and this thickness decreases outward at a rate depending upon the relative growth rates in the lateral and vertical directions. Once frazil begins to form in the flow, the growth rate of the border ice can increase substantially by the "buttering" of its edges by frazil on the river surface. This mechanism of border ice growth is particularly effective where velocities are low. However, when the concentration of surface ice becomes appreciable, there is a marked reduction in the growth of the border ice because the moving floes tend to wear away the stationary ice.

The mechanisms which control the growth of border ice by an attachment or buttering are mostly speculative. Air temperature plays a major role because the pans must freeze to the existing border ice. Similarly the local surface velocity is significant. The state-of-the- art in predicting the growth rates of border ice is not well advanced. However, Newbury (1968), Matousek (1984), and Rivard, Michel and Fonseca (1982) all have developed techniques to calculate border ice growth. These involve calculating the width of the border ice on the basis of the flow velocity at the section, the water temperature, and the heat flux. The latter term can be evaluated by either considering the air temperature and the wind, or simply employing the number of degree-days of freezing.

Andres (1988) developed an empirical approach using Newbury's data from the Nelson River. This river is similar in width and flow velocity to the large regulated rivers in Alberta and therefore can be used as an analogue for the Peace River. For rivers with a mean velocity less than 1.6 m/s, Newbury's observations suggest that the growth rate of border ice increases with the heat transfer from the water surface as represented by Equation **[7]**. The growth rate decreases as both the mean channel velocity, **V** and the ratio of the width of the border ice to the width of channel, **B/W** increase. Thus

$$\frac{dB}{dH_t} = 0.08\,V^{-3.37}(1-B/W)^{2.2} \qquad [7]$$

where all the units are in SI.

Figure 11 illustrates that the theoretical model reproduces the observed results relatively well, certainly within the accuracy expected for a process that can be dynamic and exhibit substantial variability. Upstream of Peace River, the model predicts the growth rate to within 30% of what was observed, over a period of about 60 to 90 days. Given that the border ice growth was only 15% of the total width, the accuracy of the model is more than adequate for modelling the progression of the ice cover.

4.2.2 Frazil Generation

The generation of frazil is by far the dominant ice producing mechanism. As such it has the major impact on the characteristics of the subsequent ice cover. For rivers in which the turbulence is sufficient to prevent thermal gradients from being established, and the turbulence is also of sufficient intensity to entrain ice formed in supercooled region at the water surface, the whole mass of water cools at the same rate without forming a static surface cover. The rate of cooling is proportioned to the rate of heat loss from the surface of the water and can be calculated by Equation [6]. When the water temperature reaches 0°C and heat loss continues, supercooling occurs. Through the process of nucleation, small ice crystals (frazil) form throughout the supercooled flow. These ice crystals coalesce into buoyant, loose accumulations (slush), rise to the surface, and form a crust on contact with the cold air. These masses of frazil take on a round pancake-like appearance, with a solid flat surface and a mass of porous frazil slush suspended underneath.

The above process is best visualized on a regulated river. To simplify the analyses, without losing the essence of the process, steady conditions will be assumed and the diffusion terms in the heat transfer equation will be neglected. As a parcel of water leaves the tailrace it is cooled and the water temperature, **T** at any distance, **x** downstream of the dam can be determined from

[8] $$\frac{dT}{dx} = \left[\frac{W}{\rho C_p Q}\right]\left[(1-\alpha_w)\Gamma H + h_w (T_a - T)\right]$$

where **W** is the water surface width, **Q** is the discharge, ρ is the density of water and C_p is the specific heat of water. Downstream of the nucleation point which is very close to the location where the water temperature becomes zero, the ice discharge, Q_i at any location **x** can be determined, as a first approximation by

[9] $$\frac{dQ_i}{dx} = \frac{W}{\rho_i L}\left[(1-\alpha_w)\Gamma H + h_w (T_a - T)\right]$$

where **L** is the latent heat of ice and ρ_i is the density of ice. However, as the concentration of ice increases and ice floes form, there is a significant reduction in the area of open water available for the generation of new ice and the rate of frazil production decreases in a downstream direction.

This reduction in the formation of new ice is somewhat offset by the formation of thermal ice by the thickening of the solid ice crust which caps the floes.

The reduced rate of ice generation due to the presence of floating ice at some concentration, C_i has been investigated by Hausser, Saucet, and Parkinson (1984) and Rivard et al. (1986). Their techniques require estimates of porosity of the frazil floes, the thickness of these flows, and the fraction of frazil which has integrated into the actual floe, For example, Hausser et al. (1984) assume that the thickness of the ice flows is determined by what can be achieved by thermal ice growth during the time interval that an ice floe has spent in the reach of interest. This is somewhat misleading because the thickness of the ice floes are a function more of the mean flow depth and the occurrences of rapids than the air temperature. Rivard et al. (1986) do not explicitly define an appropriate floe thickness but suggest a thickness of 0.15 m. This figure may be typical for small shallow streams, but for large rivers the thicknesses are substantially greater. The thickness of the attached slush on the North Saskatchewan River has been observed by the author to be in the order of 0.30 - 0.50 m. Rivard et al. (1986) also recommend that the porosity of the ice floes should be in the order of 0.50 and that approximately 16% of the frazil present is not integrated into the floes.

In any case, the surface ice discharge can be represented by

[10] $$\frac{dQ_i}{dx} = [1-C_i]\frac{W}{\rho_i L}[(1-\alpha_w)\Gamma H + h_w(T_a - T)]$$

and

[11] $$Q_i = WV(1-p_f)h_f C_i$$

where **V** is the mean velocity, p_f is the porosity of the frazil attached to the floes, and h_f is the thickness of the ice floes. The porosity of the ice attached to the floe probably ranges between 0.5 and 0.8 and should be greater than that of a stable ice pack because the slush is not subjected to the same consolidation forces as in an ice pack.

While the pans are travelling downstream, heat loss from the surface of the pan also produces additional ice as the liquid water within a floe freezes by heat transfer through the ice. To include the thermal ice growth phenomenon in the ice production calculation is extremely complicated because of major non-linearites arising out of the solution. Fortunately, the amount of ice generated in this mode is relatively small due to the low thermal conductivity of ice and the less efficient heat flux from the surface of the ice floe. Furthermore, this ice does not change the volume of the floe. Rather, it decreases the porosity by a slight amount.

When frazil generation begins, the water column becomes full of ice crystals (frazil). As mentioned earlier, the ice particles coalesce into buoyant accumulations of slush and rise to the surface. The rate at which the surface becomes covered with ice depends on the rate at which the frazil can coalesce into large enough accumulations with a porosity that is low enough to allow individual aggregations to rise to the surface.

This process is not well defined, but it is apparent that the rate of coalescence is a function of the size of the ice mass, its velocity, and the ambient frazil concentration. The ultimate size of the mass depends on how long it is submerged within the water column. The time of submergence is a function of the turbulence characteristics of the river which affects both the path that a mass follows on its way to the surface and the volume of the mass. If the ice mass becomes too large the turbulence can rip it apart.

It is possible to quantify the processes described above, however the final equations which describes the volume of the ice mass that might appear on the surface, and the rate at which the ice masses appear, are full of coefficients that must be evaluated by calibration. Furthermore, the resulting equation, which is a combination of the ice generation equation and the ice coalescence equation is extremely non-linear. Andres (1993) took a somewhat more practical approach, borrowing concepts from the suspended sediment technology, to solve the problem. It will be briefly described herein.

Consider an elemental volume of water with a small width **W** and length **Δx**, and depth **d** at a temperature of 0°C, moving along at a velocity **V**. This parcel of water is losing heat at a rate of $\mathbf{H_t}$ **Δx W** (1 - $\mathbf{C_i}$) and generating frazil so that the concentration of ice in the parcel is $\mathbf{C_f}$. Furthermore, frazil is being taken out of suspension at a rate **F** defined as

[12] $$F = \beta C_f$$

where the coefficient β reflects the vertical transport velocity. This equation describes the rate at which surface ice is generated.

Written in a differential form, the steady-state conservation of both surface ice and suspended ice is given in Equations **[13]** and **[14]**, respectively.

[13] $$\frac{dC_i}{dx} - \frac{\beta}{V(1-p_f)h_f} C_f = 0$$

[14] $$\frac{dC_f}{dx} + \frac{\beta}{dV} C_f - \frac{H_t}{dV \rho_i L}(1 - C_i) = 0$$

Combining Equations [13] and [14] to solve for C_i produces the following second order non-homogenous equation

[15] $$\frac{d^2 C_i}{dx^2} + \frac{\beta}{dV}\frac{dC_i}{dx} - \frac{H_t \beta}{\rho_i L(1 - p_f) h_f d V^2}(1 - C_i) = 0$$

which can be solved analytically.

This equation contains three unknown parameters: β and h_f, and p_f. The other parameters are known from the hydraulic and meteorological characteristics. Both β and h_f affect the rate of ice production and the rate at which surface ice is generated. Their values must be determined from calibration, and are specific to a given hydraulic condition. Once β and h_f are defined for a given river, C_i can be computed from Equation [15] and reintroduced into Equation [14] to calculate the concentration of the suspended frazil. Thus, the surface ice discharge and the suspended ice discharges are given as

[16] $$Q_i = WV(1 - p_f) h_f C_i$$

and

[17] $$Q_f = QC_f$$

The porosity of the floes, p_f can be determined "a priori" within a reasonable range based on observations on other rivers. If the longitudinal ice concentration gradients can be measured, then the appropriate values of β and h_f can be determined by fitting Equation [15] to the measured data. This equation is a second order homogeneous equation with constant coefficients and can be solved analytically as long as two boundary conditions are defined. These are

[18] $$C_i(0) = 0$$

and

[19] $$\frac{dC_i(0)}{dx} = 0$$

This produces a solution such that

[20] $$C_i(x) = f(\beta, h_f)$$

for any given set of meteorologic and hydraulic conditions.

The challenge is to determine the appropriate values of β and h_f that best represent the physical conditions, taking note that there are two parameters which can be adjusted and only one equation. Fortunately both β and h_f produce different attributes in the shape of the curve. The relationship between C_i and x is an S-shaped curve, which is typical of many processes where the rate of growth of a particular quantity is a function of the value of that quantity. The slope of the curve at its inflection point is a function of h_f. Small values of h_f result in a rapid increase in surface ice concentration because the thin floes rapidly take up all the available surface area. On the other hand, if the flows are assumed to be thick (large values of h_f) then the increase in concentration will be small because all the ice is used to thicken the floes rather than make more floes.

The value of β determines the curvature of the relationship immediately after surface ice beings to appear. Larger values of β, indicative of very rapid transport of ice to the surface produces a rapid growth in the ice concentration. Smaller values of β produce less rapid growth in the ice concentration and a flatter curve.

The field program to measure the longitudinal gradients in the surface ice concentration on the Peace River is described by Andres (1993). Four separate aerial observation flights were made at four different times during the year to photograph the development of the surface ice along the Peace River. Figure 12 illustrates the range of values of both h_f and β that can be prescribed to fit one set of particularly good data - that from January 27, 1992. Values of h_f between 0.1 and 0.50 where tested, in conjunction with values of β between 0.000008 and 0.00010. It is evident that values of h_f between 0.35 to 0.45 m and values of β of 0.00002 m/s are the most appropriate.

With these values, the theoretical approach slightly underpredicts the development of the surface ice at the time when it first begins to form but it adequately reflects the measured data after about 100 km of river when the concentration of floes exceeds 20%.

The one drawback to this analysis, however, is the necessity to assume a constant flow thickness, h_f. Although not entirely unreasonable, it would be more appropriate to allow for some variability in the thickness of the floe, related to the size of the flow and/or the surface concentration. In other words, the larger concentration, the greater the floe size; and hence the thicker the floe. One could arbitrarily assume a relationship between C_i and h_f, but this makes Equation [15] highly non-linear and difficult to solve either analytically or numerically. However it could explain the poor representation of the data in the first 100 km or when the concentration of the ice is less than 20%.

4.2.3 Cover Formation

On all but very slow-moving streams, the initial ice cover is formed by frazil pans and slush lodging against a solid ice cover formed by surface growth at special low-velocity locations such as at deep pools, the entrance to a lake, or the head of a reservoir. Lodgement or bridging may also occur when frazil pans lodge at a surface contraction in a long narrow reach, or where shorefast ice has grown outward from the bank, usually in a sharp bend. The ice cover then progresses upstream and an "accumulation" type of cover forms, solidifying and thickening by a variety of processes, depending upon the hydraulic and meteorologic conditions. Determining "a priori" where bridging may occur is very difficult because a rational theory for defining the hydraulic criteria for bridging has yet to be developed.

Once lodgment has occurred, two main stability criteria must be met for a stable accumulation to form and progress upstream: entrainment and internal strength. If a stable lodgement forms and the velocity is low enough that entrainment of floes does not occur, then the cover initially accumulates by the juxtaposition of floes one layer thick. On most large Alberta streams the frazil pans are sufficiently mature to form large rafts of sufficient size and integrity to resist entrainment at the head of the cover (Figure 13). Instead the solid crusts of the floes and rafts juxtapose against each other (Figure 14), and advance upstream at a rate determined by the stream velocity and surface concentration of ice. The loose frazil attached to the floe is removed from the crust by the shear of the flow and redistributed downstream under the previously formed cover. As soon as the juxtaposed cover begins to form, its strength begins to increase due to downward freezing through the frazil-filled voids between the pans. The thickness of the frozen frazil increases with the length of time the juxtaposed cover remains in place. When added to the internal strength of the thin frazil accumulation, this frozen frazil can significantly increase the strength of the accumulation.

As the ice cover advances upstream by juxtaposition of floes and the redistribution of the frazil, the downstream component of the weight of the cover and the drag on its under surface increase proportionately to its length. These forces are resisted by the frozen cover downstream and to a lesser extend by shear along the bank. When, at any location, these forces exceed the combined internal strength of the cover produced by freezing at the surface, internal friction, and cohesion, then the cover shoves and consolidates (Figure 14). The thickness increases to develop greater internal friction forces which can be transmitted, to a much greater extent, to the river banks.

Because this process normally occurs abruptly, the thickened accumulation is not significantly affected by freezing and develops an equilibrium thickness based on the cohesion and internal friction of the thickened cover, if sufficient ice is available. This form of cover can be analyzed by the stability equation proposed by Pariset, Hausser, and Gagnon (1966), from which the thickness, **h** can be computed

$$\mu \rho_i [1-(\rho_i/\rho)] g h^2] = W(\rho g R_i S + \rho_i g h S) \qquad [21]$$

where **W** is the river width; **S** is the representative uniform water surface slope; **μ** is a dimensionless coefficient that depends upon the internal friction and porosity of the jam (taken as a granular medium); $\rho g R_i S$ and $\rho_i g h S$ represent, respectively, the shear exerted by the flow on the underside of the accumulation and the downstream component of the accumulation's weight per unit area; ρ and ρ_i are the density of water and ice, respectively; and R_i is the hydraulic radius associated with the ice cover.

Thus, the stability of the ice cover at any location downstream of the ice front is a function of the relative magnitudes of the force on the cover which is given by

$$F = L_c g S(\rho R_i + \rho_i h) \qquad [22]$$

where L_c is the effective length of the ice cover (ie. that part of the ice cover that can transmit the shear stress and weight downstream to the critical location) and R_i and is given as

$$R_i = \frac{Y}{2}\left(\frac{n_i}{n}\right)^{3/2} \qquad [23]$$

where **Y** is the total flow depth under the ice cover and n_i and **n** are the Manning's roughness coefficients of the ice and the composite channel, respectively; and of the internal strength of the ice cover which is given by

$$R = \mu \rho_i (1-\rho_i/\rho) g h_j^2 + \sigma_y \tau \qquad [24]$$

where the first term represents the internal friction strength of the juxtaposed frazil accumulation.

The second term is the strength gained from downward freezing of the juxtaposed cover, where σ_y is the strength of solid ice and τ is thickness of the solid ice growing in the frazil between the floes in the juxtaposed ice cover. It is not clear how the ice cover fails, ie. by bending or crushing, however regardless of the failure mode the strength of the ice in the cover is important in determining the integrity of the newly formed ice cover.

4.2.4 Stability Criterion

The characteristics of an ice cover in any particular reach of the river depend on the rate at which ice arrives at the ice front (thereby determining the rate at which the ice cover is progressing upstream and the rate at which the force on the cover is increasing) and the rate at which the strength of the ice cover is increasing due to the growth of the intersticial ice between the juxtaposed floes. The former term depends on the the air temperature as it affects the generation of the frazil and the subsequent supply of ice arriving at the ice front; the channel width and slope, because it determines both the drag on the underside of the ice and the streamwise component of the weight of the cover; and the discharge, because it controls how these effects are manifested. The latter term depends on the thickness of the juxtaposed ice, because it provides the initial strength of the cover prior to the growth of the thermal ice; and the air temperature, because it determines the rate of growth of the thermal ice.

The sinuosity of the channel also comes into play, because it determines the limiting or maximum length of ice cover that can affect the forces at any particular location. If the channel is perfectly straight, no bank support is provided and the force at any location is simply determined from the length of cover upstream. However, if a channel is highly sinuous, then as the ice cover progresses around a bend, or for a length that is equivalent to half the meander length, then much of the force generated from the ice upstream is transmitted to the river bank at the downstream end of the bend. This insulates the ice located downstream from additional forces as the ice cover continues its progression upstream. This provides for an upper limit to how much force can be exerted on the ice cover at any particular location.

At any location x_f, the force balance on the ice cover is given by Equations [22] and [24]. Furthermore, as long as the time integral of the rate of change of the resisting force, given as

[25]
$$R(x_f) = \mu \rho_i (1-\rho_i/\rho) g h_j^2 + \sigma_y \int d\tau$$

where τ the thickness of the thermal ice, is greater than the time integral of the rate of change of the driving force, given as

[26] $$F(x_f) = gS(\rho R_i + \rho_i h) \int dx$$

then the juxtaposed cover should be stable and no thickening should occur. It is noted that the newly formed juxtaposed ice cover is thin and derives very little support from the banks because of the fluctuating water levels. Equating Equations **[25]** and **[26]**, writing **dx** as

[27] $$dx = \frac{Q_i}{Wh_f(1-p_f)}\,dt$$

and approximating **dτ** as

[28] $$d\tau = \frac{k_i T_a}{\rho_i L(1-p_f)}\,\frac{dt}{\tau}$$

where **k$_i$** is the thermal conductivity of ice, results in the stability equation that determines whether or not consolidation will occur. That is, if the following inequality is satisfied, then the juxtaposed cover should be stable.

[29] $$\mu \rho_i(1-\rho_i/\rho) g h_j^2 + \sigma_y \frac{k_i T_a}{\rho_i L(1-p_f)} \int \frac{dt}{\tau} > \frac{gS(\rho R_i + \rho_i h)Q_i}{Wh_f(1-p_f)} \int dt$$

Equation **[29]** must be integrated from **t**=0 to **t=t$_{cr}$**, where **t$_{cr}$** is the time that it takes for the ice cover to progress upstream for a length **L$_{cr}$** equal to about one quarter of a meander length. The value of **t$_{cr}$** is given as

[30] $$t_{cr} = \frac{L_{cr} W h_f(1-p_f)}{Q_i}$$

where the limiting ice discharge **Q$_i$** is given by

[31] $$Q_i = C\,W h_f (1-p_f)\,V$$

It should be noted that the actual ice discharge is a function of the length of the reach in which ice is being produced and the air temperature. It is very difficult to represent the ice discharge by a simple expression because it is best determined numerically. On the other hand, the limiting value as shown in Equation [31] is indicative of the maximum ice discharge and probably is quite representative of the ice discharge for very cold conditions with the head located well downstream of the point of intial ice production. Inserting Equations [30] and [31] into Equation [29], and carrying out the integration, results in a criteria for determining the stability of a juxtaposed cover. This is given as

[32] $$\frac{\sigma_y}{[g(\rho R_i + \rho_i h_f)S]} > [\frac{\rho_i (1-p_f) L V L_{cr}}{2 k_i T_a}]^{1/2}$$

Using the Manning formula to represent the relationship between discharge and the flow depth and velocity and writing V and R_i in terms of the discharge arriving at the ice front and the discharge under the ice cover, respectively; and simplifying by assuming that the weight of the thin juxtaposed ice cover is about equal to the resisting shear due to friction along the banks, and both are much less than the drag on the bottom of the ice cover, ie. $h_f < R_i$; and that the roughness of the underside of the ice is similar to that of the stream bed, results in a stability equation given as

[33] $$[\frac{L_{cr}}{T_a}]\,[\frac{Q}{W}]^{8/5} S^{17/10} < \frac{7.4\,\sigma_y^2 k_i}{\rho^2 g^2 n^{3/5} L \rho_i (1-p_f)}$$

The left side of the inequality represents the effects of the discharge in terms of its contribution to the rate at which ice arrives at the ice front and the shear force under the juxtaposed ice cover, the air temperature as it relates to the growth of the solid ice within the juxtaposed cover, the slope which affects the shear stress, and the critical reach length over which the stability must be maintained. The right side length of the inequality represents a numerical constant which results from the way in which the Manning equation represents flow under the ice cover, the strength of the newly formed ice between the ice floes, and a number of constants related to the roughness of the stream and the underside of the ice, the ice characteristics, and the growth mechanism of the solid ice. The right side of the equation can be evaluated theoretically to be in the order of 0.0010 to 0.010 for typical values of composite roughness, floe porosity, and ice strength. As long as the inequality holds, the juxtaposed form of the ice cover should dominate.

4.3 FREEZE-UP ON A NON-REGULATED RIVER

Freeze-up on a non-regulated river follows many of the same physical principles (heat loss, border ice growth, frazil generation, lodgement, and stable cover formation) as for a regulated river, except that temporal gradients are more important than spatial gradients and different processes dominate the mechanisms by which a stable cover forms. Whereas under the regulated condition, streamwise gradients in water temperature and ice concentration are established, and a relatively orderly upstream progression of the ice cover follows under a more or less constant discharge regime; in the non-regulated condition there are no streamwise gradients in temperature or ice, the discharge is usually low and receeding, and in any given reach the ice cover forms at a variety of locations almost at the same time. The entire river cools at essentially the same rate, depending on the air temperature and the flow depth, with the only warming influence being the groundwater inflow.

Extensive border ice can develop because of the low velocity and flowing ice usually will appear everywhere at about the same time. The development of the first elements of the ice cover is a function of the decrease in flow as it relates to the decrease in the flow width and a corresponding increase in the width variablity. This promotes multiple lodgements, that usually provide for a number of short localized areas with a stable ice cover. Freeze-up occurs mainly by the juxtaposition of ice floes formed from the locally generated frazil. Shoving may occur now and then but it is not an important process in developing a stable ice cover unless the river has an extremely steep slope. The stage increase associated with freeze-up is small and due mainly to the additional roughness of a rather thin ice cover. Furthermore, the stage increase is offset by the reduction in discharge following the recession curve into the winter period.

Unfortunately there are not a lot of observations of freeze-up under these conditions and the thickness of the ice cover immediately at freeze-up is one of conjecture. If some estimates of the initial thickness of the cover can be made then it is a relatively simple procedure to calculate time of occurence and the freeze-up water level increase for the non-regulated condition.

4.3.1 Water Temperature

Thw whole mathematical concept of freeze-up on a non-regulated river hinges on the assumption of the existance of individual reaches into which there is no ice inflow or from which there is no ice outflow. As the length of the reach increases, this assumption becomes more valid, and on the scale of 100 km, it probably is quite accurate. Because there is no single dominant heat source in the system, there are no spatial gradients to deal with and the temporal gradients dominate. This is in contrast to the regulated condition where the spatial gradients at a particular location are much greater than temporal gradients of the water temperature.

The rate at which the water in the reach cools depends on the air temperature, discharge, and the depth of flow in the reach. A simple energy balance equation can be used to calcualate this rate.

[34] $$\frac{dT}{dt} = \frac{1}{\rho C_p d}[h_w(T_a - T) + (1-\alpha_w)\Gamma H]$$

All the parameters in the above equation have been defined previously. Assuming a more or less constant flow depth **d** (due to a more or less constant discharge), and some average air temperature that persists over the time when the water temperature declines from some initial temperature $\mathbf{T_o}$ to near zero, the above equation can be solved to represent the temporal variation in the water temperature. In reality, one would probably solve the equation numerically using a finite difference scheme to allow for the representation of the changing air temperature and discharge for each particular year of concern. The finite difference equation is as follows

[35] $$T_{i+1} = T_i + [h_w(T_{ai} - T_i) + (1-\alpha_w)\Gamma H]\frac{\Delta t}{\rho C_p d_i}$$

where

[36] $$d_i = [\frac{Q_i n_b}{W S^{1/2}}]^{3/5}$$

and $\mathbf{n_b}$ is Manning's bed roughness. Equation [35] suggests that the cooling of the water follows very closely the change in the air temperature in the basin.

4.3.2 Ice Cover Formation

When the water temperature reaches the freezing point, frazil ice is generated in the bulk of the flow and because of its buoyancy it rises to the surface and forms ice floes. These ice flows are similar to those formed on the regulated river and there is no reason not to assume similar values of floe thickness and porosity that was defined in the previous sections. These ice floes are the main building blocks of the initial ice cover that forms following the production of the frazil. This process can be described mathematically as follows.

1. Assume a control reach of length $\mathbf{L_r}$ with water surface width of **W**.

2. The frazil that is generated forms ice floes having a given diameter and thickness, $\mathbf{h_f}$.

3. The fraction of the surface of the control reach that is taken up by the frazil pans is designated as C_i. As this value increases, the amount of area available for ice generation and subsequently the rate of frazil generation is reduced.

4. As the river becomes totally filled with ice, say at a value of C_i of 0.95, the frazil generation rate approaches zero and an equilibrium ice concentration is achieved for the ambient hydraulic characteristics (discharge, velocity, width). At this time no additional ice is generated, and if the flow remains constant, there is no mechanism to thicken the ice and cause lodgement.

5. As the discharge drops and the attendent width decreases, the ice cover thickens due to the overall reduction in the surface area of the river. This promotes thickening and when the ice cover is sufficiently thick to develop friction along the bank and reduce the ice velocity, thermal freezing produces lodgement and a stable ice cover develops.

The increase in the volume of ice produced and the surface ice concentration can be can be described by the following equations

[37]
$$\frac{dV_i}{dt} = \frac{WL_r h_w T_a}{\rho_i L}(1-C_i)$$

[38]
$$C_i = \frac{V_i}{h_f W L_r (1-p_f)}$$

where V_i is the volume of ice formed. When Equations [37] and [38] are combined and solved, they give an expression for the fraction of ice cover in a particular reach of the river at any given time after the start of the generation of the frazil, t_o. It should be noted that it is not necessary to include the solar radiation component in this equation. By late October the sun angle is relatively low and the hours of bright sunshine are limited to such an extent that input of solar radiation, in terms of the total heat budget, is very small. The expression for the ice concentration is

[39]
$$C_i = 1 - \exp\left[\frac{h_w T_a (t-t_o)}{h_f \rho_i L (1-p_f)}\right]$$

The choice of the limiting C_i, after which no additional ice is generated probably has to be determined through some form of calibration. A value of 0.95 may be appropriate. Furthermore, as mentioned earlier, the appropriate value of $\mathbf{h_r}$ also is not explicitly defined, but has been assumed from observations for regulated conditions. As for Equation **[39]**, it may be more appropriate to write the solution of the equation in a finite difference scheme to allow for variable air temperatures from day to day.

The stage decrease that is required to produce lodgement is quite situational and needs to be verified from observations. This will ultimately determine the time of freeze-up. The stage increase associated with this type of ice cover in a given reach can be calculated from conventional hydraulics as long as the thickness and roughness of the accumulation is known, and the rate of discharge recession can be estimated.

5.0 FREEZE-UP OBSERVATIONS

Observations of ice conditions on the Peace River did not begin in any formal manner until after regulation. Some concerns regarding the effects of regulation were identified in the Peace/Athabasca Delta Study, however the effects on the water levels in the Delta were not specifically related to the ice regime of the upper part of the river. As early as 1973, high water during breakup at Peace River town resulted in concerns about the operation of the dam and the role that ice had in producing high water conditions. This precipited an extensive ice observation program by BC Hydro and a number of studies of breakup jams on both the Peace and Smoky Rivers in the vicinity of Peace River were undertaken. Attention to the the freeze-up process did not occur until the late 1970's when very high water levels at Peace River town accompanied the annual freeze-up.

Since the late 1970's, AEP has observed the progression of the ice cover over the entire winter to assist in the mitigation of high water levels caused during freeze-up and breakup at Peace River town. The management philosophy focussed on producing freeze-up at as high a steady discharge as possible to ensure that the cover formed as thickly as possible. This would maximize the cover strength to withstand discharge fluctuations during the remaining winter. Under this operating regime, dangerously high water levels resulted on three different occasions as part of the normal freeze-up process under the regulated conditions. It should be noted that apart from some detailed work undertaken by Acres in 1982 on behalf of the Canadian Electrical Association, and a couple of studies undertaken by ARC in the 1980's, very little work has focussed on observing and quantifying freeze-up and breakup processes by any of the agencies involved in the management of the river.

In the 1990's, interest in the ice regime increased due to the MBIS by Environment Canada and the NRBS undertaken jointly by Canada, Alberta, and the Northwest Territories. More detailed analyses of many of the freeze-up processes were undertaken for the development of a model to predict the ice cover progression rates to assess the potential impacts of climate change on the ice regime. This allowed for the evaluation of some of the current theories of freeze-up, and resulted in establishing a theoretical framework for the numerous ice observations undertaken by AEP. Furthermore, a recent winter dispersion study on the river provided an opportunity to characterize freeze-up on a reach scale (rather than at a gauge site, for example) and to actually determine the hydraulic characteristics of the flow under the ice cover.

Most of the above work did very little to clarify the ice regime downstream of Fort Vermilion. There was a significant gap in knowledge related to how the ice cover formed downstream of the Vermilion Chutes and whether the ice cover staged up over the Chutes, or there was independent bridging upstream of the Chutes. This current study has allowed for some observations to be taken downstream of Fort Vermilion for the 1993 freeze-up. This has resulted in a better understanding of the dominant processes and a definition of the downstream boundary conditions for any numerical modelling which might be attempted for the river as a whole.

In addition to these miscellaneous studies, WSC has operated gauges through the study area since at least 1958. These gauges, located at Taylor, Peace River, and Peace Point, provide site-specific information on water temperatures, discharges, freeze-up dates, freeze-up stages, and also similar information related to breakup. These records alone do not allow for a characterization of the ice processes on the entire river. However, when used with a numerical model or as a form of calibration for some process-based approach for quantifying the ice regime, the data becomes very useful. Tables 4, 5, and 6 summarize the available infromation from the gauge records. It should be noted that the in many years the gauges becomes inoperative during freeze-up and no data, or only estimates of water levels following the development of an ice cover, are available.

The information in the following tables has been determined from the interpretation of the strip charts and the advanced records. From these records, the air temperature and discharge at freeze-up, the pre-freeze-up stage, the maximum instantaneous ice-related stage, and the more-or-less constant post-freeze-up stage are identified. The intent of the tables is to present the discharge just prior to freeze-up before any backwater effects destroyed the reliability of the discharge estimate. The maximum stage usually is associated with a surge of water resulting from a consolidation. This peak is more a reflection of the short term increase in discharge, which itself is difficult to quantify and is not really related to the thickness of the stable ice cover. The stage after the ice cover forms is chosen to reflect the stage that is associated with the freeze-up discharge, rather than the discharge after freeze-up. This latter flow is somewhat reduced because of storage under the ice cover upstream for the regulated condition or because of reduced flows due to the normal recession of the hydrograph for the natural or non-regulated condition.

The interpretation of the gauge records is more difficult for the natural conditions or for regulated conditions during which the discharge is low and hence there is no dramatic thickening of the ice cover, or when the slope and the velocity are low enough such that juxtapositon is the dominant form of ice cover development regardless of the discharge. Figure 15 illustrates the difference in the response of the stage to both the types of freeze-up and also identifies the parameters that are summarized in the tables.

Figures 16 and 17 illustrate the global effects of regulation on the timing of freeze-up and the stage increase associated with the freeze-up event. These figures were constructed from the data in Tables 4, 5, and 6 and therefore reflect the characteristics of the ice regime only at the three gauge locations. It should also be noted that the quality of the data is generally lower prior to regulation and even after regulation there are significant gaps in the data. Taylor has no ice-related information prior to regulation while there are some difficulties with the Peace Point data after regulation.

Generally, the figures confirm the current thoughts on the effects of regulation. The timing of freeze-up has been delayed the most at Taylor. It is known that an ice cover did form early in the winter prior to regulation. After regulation, no stable ice cover forms, or if it does it usually occurs after Feb 1. At Peace River, the dominant date of freeze-up has been shifted by at least a month and the distribution is skewed towards the late part of the winter. At Peace Point, there is not a dramatic shift in the freeze-up date after regulation. This suggests that the temperature regime at this location has not changed due to the addition of a heat source at the Bennett Dam and the increased flows. The location is sufficiently far downstream of the dam so that the normal temperature regime is uneffected by the warmer outflow from the Dam.

Table 4 Summary of freeze-up dates, discharges, and water levels at Taylor

WSC gauge:		Peace River near Taylor, 07FD002													
AES station:		Fort St. John A													
Year	Date of First Ice Effect	Prefreeze-up Characteristics				Freeze-up		Maximum Freeze-up Level				Average Freezeup Level			
		Representative Date	Discharge (m^3/s)	Gauge Height (m)	Elevation (m)	Date	Mean Daily Air Temperature (°c)	Date	Gauge Height (m)	Elevation (m)	Stage Increase (m)	Representative Date	Gauge Height (m)	Elevation (m)	Stage Increase (m)
1958-59	Nov-16	Nov-15	660			Nov 16-30									<0.5
1959-60	Nov-15	Nov-14	538	5.355		Nov 1-15									<0.5
1960-61	Nov-25	Nov-24	575	1.393	401.666	Nov 16-30									<0.5
1961-62	Nov-29	Nov-28	428	1.000	401.273	Nov 16-30									<0.5
1962-63	Dec-09	Dec-08	561	1.362	401.635	Dec 1-15									<0.5
1963-64	Nov-19	Nov-18	442	0.899	401.172	Nov 16-30									<0.5
1964-65	Nov-18	Nov-17	985	1.734	402.007	Nov 16-30									<0.5
1965-66	Nov-17	Nov-16	844	1.426	401.699	Nov 16-30									<0.5
1966-67	Nov-17	Nov-16	657	0.817	401.090	Nov 16-30									<0.5
1967-68	Dec-01	Nov-30	433	0.829	401.102	Dec 1-15									<0.5
1968-69	-					Dec 1-15									<0.5
1969-70	-					-									
1970-71	Jan-01	Dec-31	1102	1.942	402.215	>Feb 15									1.5-2.0
1971-72	Jan-13	Jan-12	1161	2.024	402.296	Jan-14	-26.9	Jan-15	5.934	406.207	3.911	Jan-16	5.928	406.201	3.904
1972-73	Dec-30	Dec-29	1546	2.643	402.915	>Feb 15									
1973-74	Jan-01	Jan-14	1011	2.640	402.912	Jan-15	-28.6	Jan-16	4.935	405.207	2.295	Jan-17	4.822	405.094	2.182
1974-75	Feb-02	Feb-01	1506	2.637	402.909	>Feb 15									
1975-76	None					>Feb 15									
1976-77	None					>Feb 15									
1977-78	None					>Feb 15									
1978-79	Feb-15	Feb-14	1740	2.766	403.039	Feb-15	-31.3	Feb-18	6.550	406.823	3.784	Feb-19	6.181	406.454	3.415
1979-80	None					>Feb 15									
1980-81	None					>Feb 15									
1981-82	None					>Feb 15									
1982-83	Jan-13	Feb-04	480	1.732	402.005	Feb-06	-20.0	Feb-10	4.610	404.883	2.878	Feb-11	4.487	404.760	2.755
1983-84	Jan-04	Jan-03	1610	2.668	402.941	>Feb 15									
1984-85	Jan-01	Feb-09	1220	2.309	402.582	Feb-10	-28.5	Feb-12	6.049	406.322	3.740	Feb-13	5.913	406.186	3.604
1985-86	None					>Feb 15									
1986-87	None					>Feb 15									
1987-88	None					>Feb 15									
1988-89	Jan-24	Feb-02	1540	2.706	402.979	>Feb 15									
1989-90	None					>Feb 15									
1990-91	None					>Feb 15									
1991-92	-	Feb-21	1820	3.048	403.321	Feb-24	-	Feb-24	5.099	405.372	2.051	Feb-24	5.099	4.05.372	2.051
1992-93	Nov-17	Nov-16	1730	2.941	403.214	>Feb 15									
-	No Data														

Table 5 Summary of freeze-up dates, discharges, and water levels at Peace River

WSC gauge		Peace River at Peace River, 07HA001														
AES station:		Peace River A														
Year	Date of First Ice Effect	Prefreeze-up Characteristics				Freeze-up		Maximum Freeze-up Level				Average Freeze-up Level			
		Representative Date	Discharge (m3/s)	Gauge Height (m)	Elevation (m)	Date	Mean Daily Air Temperature (°c)	Date	Gauge Height (m)	Elevation (m)	Stage Increase (m)	Representative Date	Gauge Height (m)	Elevaton (m)	Stage Increase (m)
1958-59	Nov-08	Nov-22	334	7.544	312.344	Nov-28	-	Dec-01	8.141	312.941	0.597	Dec-02	8.102	312.902	0.558
1959-60	Nov-13	Dec-16	572	8.031	312.831	Dec-18	-10.0	Dec-20	9.333	314.133	1.302	Dec-21	9.147	313.947	1.116
1960-61	Nov-08	Dec-03	391	7.507	312.307	Dec-04	-20.8	Dec-11	8.147	312.947	0.640	Dec-12	8.132	312.932	0.625
1961-62	Oct-24					Oct 15-30									0.5-1.0
1962-63	Nov-14	Dec-11	504	7.885	312.685	Dec-12	-5.8	Dec-13	8.519	313.319	0.634	Dec-14	8.461	313.261	0.576
1963-64	Nov-04	Nov-09	920	6.437	311.237	Nov-10	-7.2	Nov-10	7.955	312.755	1.518	Nov-11	7.815	312.615	1.378
1964-65	Nov-08	Dec-06	433	6.343	311.143	Dec-07	-4.2	Dec-09	7.666	312.466	1.323	Dec-10	7.452	312.252	1.109
1965-66	Nov-05	Dec-01	408	6.151	310.951	Dec-02	-8.9	Dec-07	6.782	311.582	0.631	Dec-08	6.770	311.570	0.619
1966-67	Nov-01	Nov-25	733	6.608	311.408	Nov-27	-23.3	Dec-01	7.047	311.847	0.439	Dec-02	7.023	311.823	0.415
1967-68	Nov-09					Nov 1-15									<0.5
1968-69	Nov-07	Nov-21	195	5.624	310.424	Nov-22	-4.4	Nov-24	6.066	310.866	0.442	Nov-23	6.041	310.841	0.417
1969-70	Nov-17	Nov-22	909	6.696	311.496	Nov-26	-7.2	Nov-28	7.516	312.316	0.820	Nov-30	7.233	312.033	0.537
1970-71	Nov-24	Nov-22	926	6.645	311.445	Dec-01	-26.1	Dec-04	7.614	312.414	0.969	Dec-05	7.513	312.313	0.868
1971-72	Nov-28	Dec-12	810	6.706	311.506	Dec-13	-27.8	Dec-17	8.336	313.136	1.630	Dec-18	8.355	313.155	1.649
1972-73	Dec-03	Dec-20	1184	7.041	311.841	Dec-22	-31.1	Dec-24	9.696	314.496	2.655	Dec-25	9.598	314.398	2.557
1973-74	Nov-06	Dec-06	1164	6.995	311.795	Dec-07	-14.7	Dec-09	9.107	313.907	2.112	Dec-10	8.912	313.712	1.917
1974-75	Jan-04	Jan-11	1240	6.992	311.792	Jan-13	-21.4	Jan-18	10.354	315.154	3.362	Jan-19	10.314	315.114	3.322
1975-76	Dec-01	Dec-09	835	6.754	311.554	Dec-12	-30.8	Dec-15	9.839	314.639	3.085	Dec-16	9.641	314.441	2.887
1976-77	Dec-06	Jan-12	1722	7.230	312.030	Jan-13	-21.2	Jan-17	9.790	314.590	2.560	Jan-18	9.778	314.578	2.548
1977-78	Dec-04	Dec-05	1960	7.324	312.124	Dec-06	-33.6	Dec-12	10.110	314.910	2.786	Dec-13	10.095	314.895	2.771
1978-79	-	Dec-30	1800	7.123	311.923	Dec-31	-	-				Jan-02	9.683	314.483	2.560
1979-80	Dec-12	Dec-21	643	6.342	311.142	Dec-22	-7.2	Dec-24	7.693	312.493	1.351	Dec-25	7.519	312.319	1.177
1980-81	Nov-30	Dec-06	1010	6.756	311.556	Dec-07	-23.5	Dec-10	9.347	314.147	2.591	Dec-11	9.122	313.922	2.366
1981-82	Jan-01	Jan-01	2010	7.425	312.225	Jan-02	-31.3	Jan-03	9.426	314.226	2.001	Jan-04	8.813	313.613	1.388
1982-83	Dec-07	-				Dec-16		-				-			2.0-2.5
1983-84	Dec-12	Dec-14	1620	7.168	311.968	Dec-17	-29.2	Dec-18	9.642	314.442	2.474	Dec-18	9.493	314.293	2.325
1984-85	Dec-19	Dec-18	1870	7.134	311.934	Dec-21	-29.2	Dec-23	10.702	315.502	3.568	Dec-24	10.617	315.417	3.483
1985-86	Dec-01	Nov-29	1830	7.210	312.010	Dec-03	-25.0	Dec-08	10.263	315.063	3.053	Dec-09	10.204	315.004	2.994
1986-87	Jan-01	Jan-14	1480	7.001	311.801	Jan-18	-1.5	Jan-25	10.955	315.755	3.954	Jan-26	10.857	315.657	3.856
1987-88	Jan-01	Jan-30	1760	7.172	311.972	Feb-01	-28.3	Feb-09	10.107	314.907	2.935	Feb-10	10.032	314.832	2.860
1988-89	Dec-29	Dec-23	1300	6.787	311.587	Jan-02	-6.0	Jan-05	10.903	315.703	4.116	Jan-06	10.691	315.491	3.904
1989-90	Dec-21	Jan-05	1510	6.973	311.773	Jan-08	-6.4	Jan-10	10.245	315.045	3.272	Jan-11	9.990	314.790	3.017
1990-91	Nov-28	Dec-18	1700	6.984	311.784	Dec-20	-35.5	Dec-24	9.785	314.585	2.801	Dec-26	9.698	314.498	2.714
1991-92	Feb-02	Feb-08	1960	7.492	312.292	Feb-11	-21.6	Feb-15	12.150	316.950	4.656	Feb-16	12.021	316.821	4.529
1992-93	Dec-29	Dec-28	1810	7.038	311.838	Dec-29	-43.1	Jan-02	9.618	314.418	2.580	Jan-03	9.590	314.390	2.552
-	No Data														

Table 6 Summary of freeze-up dates, discharges, and water levels at Peace Point

WSC gauge:		Peace River at Peace Point, 07KC001													
AES station:		Fort Chipewyan A													
Year	Date of	Prefreeze-up Characteristics				Freeze-up		Maximum Freeze-up Level				Average Freeze-up Level			
	First Ice Effect	Representative Date	Discharge (m^3/s)	Gauge Height (m)	Elevation (m)	Date	Mean Daily Air Temperature (°c)	Date	Gauge Height (m)	Elevation (m)	Stage Increase (m)	Representative Date	Gauge Height (m)	Elevation (m)	Stage Increase (m)
1958-59	-														
1959-60	Nov-05					Nov 1-15									<0.5
1960-61	Nov-01					Nov 1-15									<0.5
1961-62	Nov-03					Oct 15-30									1.0-15
1962-63	Nov-07	Nov-18	1373	20.321		Nov-20	-11.4	Nov-20	21.159		0.838	Nov-21	21.074		0.753
1963-64	Nov-15	Nov-14	963	3.426	210.547	Nov-15	-6.4	-				Nov-27	3.368	210.489	<0.5
1964-65	Nov-08					Nov 1-15									<0.5
1965-66	Nov-09	Nov-06	2285	5.410	212.531	Nov-10		Nov-11	6.821	213.942	1.411	Nov-13	6.175	213.296	0.765
1966-67	Oct-26	Oct-27	929	3.435	210.556	Nov-01	-10.3	Nov-06	5.389	212.510	1.954	Nov-07	5.319	212.440	1.884
1967-68	Nov-06	Nov-13	787	3.395	210.516	Nov-14	-6.1	Nov-19	4.374	211.495	0.979	Nov-20	4.365	211.486	0.969
1968-69	Nov-06	Nov-14	538	3.338	210.459	Nov-15	-12.8	Nov-16	3.752	210.873	0.414	Nov-17	3.587	210.708	0.250
1969-70	Oct-26	Nov-05	980	3.545	210.666	Nov-11	-15.6	Nov-14	5.992	213.113	2.447	Nov-15	5.782	212.903	2.237
1970-71	Nov-14	Nov-14	991	3.414	210.535	Nov-18	-14.2	Nov-20	5.252	212.373	1.838	Nov-21	5.142	212.263	1.728
1971-72	Oct-30					Nov 1-15						Nov-26	5.772	212.893	2.017
1972-73	Nov-01					Nov 1-15						Nov-28	6.293	213.414	1.474
1973-74	Nov-05	Nov-09	1637	4.627	211.748	Nov-10	-13.1	Nov-11	5.511	212.632	0.884	Nov-12	5.508	212.629	0.881
1974-75	Nov-18	Nov-21	1832	4.773	211.894	Nov-23	-19.7	Nov-24	6.014	213.135	1.241	Nov-25	5.989	213.110	1.216
1975-76	Oct-31	Oct-30	1379	4.078	211.199	Nov 1-15						Nov-21	6.233	213.354	2.155
1976-77	Nov-11	Nov-20	141	5.069	212.190	Nov 16-30						Dec-01	7.309	214.430	2.240
1977-78	Nov-07	Nov-16	1334	5.380	212.501	Nov-20	-19.9	Nov-21	6.876	213.997	1.496	Nov-22	6.797	213.918	1.509
1978-79	Nov-10	Nov-09	1591	4.456	211.577	Nov-18	-22.9	-				Nov-23	6.157	213.278	1.701
1979-80	Nov-13					Nov 1-15									1.0-1.5
1980-81	Nov-10	Nov-09	1710	4.650	211.771	Nov 16-30						Dec-05	4.877	211.998	0.227
1981-82	Nov-15					Nov 1-15									1.5-2.0
1982-83	Nov-01	Nov-07	1420	4.381	211.502	Nov 1-15						Nov-13	6.066	213.187	1.685
1983-84	Nov-13					Nov 1-15									<0.5
1984-85	Oct-30	Oct-30	1730	5.501	212.622	Oct-31	-23.5	Nov-02	7.934	215.055	2.433	Nov-03	7.380	214.501	1.879
1985-86	Nov-04	Nov-03	2020	5.178	212.299	Nov 1-15						Nov-17	6.570	213.691	1.392
1986-87	Oct-24					-									-
1987-88	Nov-19	Nov-15	1570	4.416	211.537	Nov-16	-	-				Nov-28	7.016	214.137	2.600
1988-89	Nov-03	Nov-11	1730	4.908	212.029	Nov-13	-12.0	Nov-17	7.297	214.418	2.389	Nov-18	7.167	214.288	2.259
1989-90	Nov-06	Nov-06	1000	3.468	210.589	Nov-09	-7.5	Nov-12	5.894	213.015	2.426	Nov-12	5.688	212.809	2.220
1990-91	Oct-27	Nov-07	1930	4.755	211.876	Nov-11	-24.0	Nov-18	7.072	214.193	2.317	Nov-19	7.009	214.130	2.254
1991-92	-	Oct-29	1850	5.152	212.273	Nov-01	-13.3	Nov-03	8.047	215.168	2.895	Nov-04	7.915	215.036	2.763
1992-93	-	Nov-20	1810	5.005	212.126	Nov-24	-8.3	Nov-27	7.542	214.663	2.537	Nov-28	7.500	214.621	2.495
-	No Data														

The stage increase that is associated with freeze-up has increased dramatically after regulation. At all three locations there has been at least a 2 to 3 m increase in the "backwater" associated with the formation of the ice cover. This is due to two main effects. First, the discharge does not decrease at the normal recession rate as it did under natural conditions prior to regulation. Thus there is no compensating effect for the increase in stage related to the development of a relatively thin ice cover and the additional boundary on the flow. Second, the the discharge at all locations along the river is much higher after regulation and this results in the formation of a much thicker ice cover. This effect is especially apparent in the steep reaches upstream of Dunvegan.

5.1 PRE-REGULATION

5.1.1 Water Temperatures

Under natural, non-regulated, conditions the water temperature in the main stem of the Peace River and in the major tributaries essentially refelected the air temperature and the solar radiation input. Figure 18 illustrates the trends in the water temperatures recorded during the discharge measurements undertaken by WSC. These measurements, taken at approximately monthly increments during the period of time between 1961 and 1975, reflect the non-regulated conditions at Peace River, Fort Vermilion, and Peace Point, event though the last three or four years of data were in the regulated period. A comparible trend is also shown for the Smoky River at Watino.

It is evident that the average (or "normal") water temperature is not dramatically different for the four locations and the two rivers. Certainly, the meteorological variables are similar for all four sites. Furthermore, it can be argued from regime considerations that the flow depth, which is the only hydraulic variable which affects the water temperaure (see Equation [35]), probably scales with the discharge and the width of the river so that the depths are comparible at all four locations even if the discharges are not. Thus it is not surprising that the water temperatures behave similarily. It is evident from the figure that the water temperature typically cools to zero degrees sometime in late October or early November, depending on the temperature regime of the particular year. This provides a rather uniform time for the annual initialization of the ice generation process prior to regulation and suggests that ice would first appear on the river in early November.

Although the scope of this report precluded evaluating large volumes of data for a wide range of individual years, Equation [35] was solved in its finite difference form for the Peace River at Peace River using daily values of the air temperature, solar radiation, and pre-regulated discharges interpolated from the normal monthly values. The results of the simulation of the water temperature is shown in Figure 19 for a range of heat transfer coefficients. It should be noted that the input of the solar radiation was determined by assuming an albedo of 0.05 and an exposure factor of 0.75. The exercise suggests that the water temperature model is appropriate for calculating the water temperatures and that an appropriate value of the heat transfer coefficient should range between 10 and 15 W/m^2-°C. Furthermore, the simulation suggests that a forecast of the time when the water temperaure approaches zero and ice begins to form on a non-regulated river can be made simply by forecasting the date at which the air temperature generally falls below freezing.

One final comment must be made regarding the inflow of groundwater. No consideration was given to the inflow of heat from groundwater. Since the groundwater inflow is generally variable and not well quantified, it would have been a rather speculative exercise. Neglecting the groundwater would produce a slight under-estimation of the heat transfer coefficient since the extra heat provided by the groundwater does not have to be removed. This would suggest that the value of the real heat transfer coefficient should tend towards the upper value of 15 W/m^2-°C. This would put it more in line with the results of other studies (Andres, 1984).

5.1.2 Ice Cover Formation

The process by which a stable ice cover forms in a natural non-regulated situation is not well understood. One can speculate with a fair degree of confidence about the formation of ice floes and growth of border ice, but there is little information available to characterize how the stable ice cover actually developes. The limited scope of this project precluded an in depth evaluation of these process, however, a cursory examination of four years of records at Peace River has shed some light on this issue.

Table 7 summarizes the natural freeze-up variables for four years for which there was adequate data. The most striking revelation from the data is the long period of time between the day at which a substantial amount of ice should have been evident in the river (characterized by a concentration of 95%) and the date at which a stable ice cover formed (as infered from the water levels). It should be noted that the accuracy of the computed date of first ice is verified to within three to ten days of the first actual observation and therefore can be taken with some degree of confidence. The ice floe generation model suggests that after the date of first ice, about ten to fifteen days of ice production are required before the amount of ice in the river becomes significant. After that time, an additional twenty to thirty days are required for a stable ice cover to form. This length of time is somewhat disconcerting, and suggests that the rate of ice production may be over estimated and/or the air temperature must be very low before a substantial amount of strength can be developed due to freezing within the moving pack.

One criterion that must be satisfied to ensure that lodgement occurs appears to be related to the decrease in the flow width as the discharge decreases. Once the concentration of ice becomes very large, and ice continues to form, the river will continue to transport ice at full capacity regardless of the discharge. As the water discharge decreases, the width decreases and the variability in the width increases (the channel width is generally more variable in space at low stage). At some critical width the ice will lodge and a stable cover will form almost instantaneously over fairly long reaches. The data suggests that, regardless of the flow in the river at the time of maximum ice concentration, the discharge must decrease by an amount which will cause the width to reduce by about ten to twenty percent in order to produce lodgement at Peace River town.

The resulting ice cover will be slightly thicker than the thickness of the floes in the river just prior to the time of lodgement and the total stage increase (assuming no decrease in the flow) will be related to the thickness of the ice cover and the existance of an additional boundary on the flow. The data suggests that this increase is in the order of about 0.5 to 1.0 m (Figure 17). However it is difficult to resolve the effects of decreasing discharge in this estimate.

Table 7 Summary of freeze-up characteristics at Peace River prior to regulation

Year	1962	1965	1966	1968
Calculated date of first ice production[1]	Nov 9	Nov 2	Oct 21	Nov 4
Date of first ice observation[2]	Nov 14	Nov 5	Nov 1	Nov 7
Characteristics at C = 95%:				
Date	Nov 19	Nov 9	Nov 5	Nov 17
Discharge (m^3/s)	1135	1420	1230	450
Average width (m)	420	440	430	360
Floe thickness[3] (m)	0.45	0.45	0.45	0.45
Characteristics at lodgement:				
Date[4]	Dec 11	Dec 2	Nov 27	Nov 22
Discharge (m^3/s)	500	410	730	200
Average width (m)	370	360	390	310
Width decrease (%)	12	18	9	13
Calculated freeze-up ice thickness (m)	0.51	0.55	0.50	0.52

[1] Determined form temperature records and assumed as the day on which the mean daily air temperature is consistantly below freezing.
[2] Determined from WSC observations.
[3] Assumed from ice floe generation calibrations.
[4] Determined from interpretation of WSC records.

5.2 REGULATED CONDITIONS

Under regulation, the river is normally ice-affected for about six months of the year, in one reach or another between the WAC Bennett Dam and the Slave River. The ice regime, and particularly the timing of freeze-up, is a function of both the climate and the discharge in the river, the latter which depends on the way Lake Williston is managed. Therefore, changes in either the temperature regime or the discharge in the river will have dramatic effects on the timing of freeze-up and ultimately on the ice regime for the entire year.

The development of the ice cover is a function of air temperature and discharge in the river. These two parameters overlay the geomorphic framework (slope, width, bed material) which determines the hydraulic characteristics of the river, to produce a typical freeze-up pattern along the river. The timing and severity of freeze-up are also affected by conditions that prevail at the boundaries of the system. These are largely independent of the processes by which the ice cover forms, but must be specified to quantify the development of the ice cover. The two main boundary conditions which need to be known are (1) the upstream water temperature of the flow at the Bennett Dam and (2) the time at which the ice cover forms downstream of the Vermilion Chutes. The former identifies the amount of heat that is entering the system and the latter defines the time of the initial lodgement.

Data provided by BC Hydro suggests that the water temperature at the Bennett Dam varies very little from year to year. Figure 20 illustrates the annual variability of the initial water temperature at the tailrace and the adopted initial water temperature used in the analysis which will be described in the following sections. It appears that there is only a maximum of a two degree variation in the water temperature on any given day, from data over nine years of records under dramatically different air temperature conditions. Also, the slope of the curve is the same from year to year. Thus, the temperature condition at the upstream boundary is unique, is a function only of the day of the year, and seems to be independent or only weakly dependent on the air temperature in the basin.

5.2.1 Downstream of Vermilion Chutes

The date of the initial ice cover formation downstream of Vermilion Chutes must be known to simulate the upstream progression of the ice cover and to forecast the timing of freeze-up at any particular location. Very little information is available for that reach of the river, short of the WSC data at Peace Point. To augment this information, the freeze-up process in the reach downstream of the Vermilion Chutes and the ice cover progression over the Chutes was observed from aircraft during the period between Nov 30 and Dec 12, 1993.

As mentioned earlier, the timing of freeze-up at Peace Point has not changed dramatically due to regulation. The river still freezes up in early to mid November, albeit the stage at freeze-up is one to two metre higher. The observations of freeze-up downstream of the Vermilion Chutes indicate that the entire lower river generally freezes in the same manner as would a large lake. That is, the slope is sufficiently low that, even with the high discharges coming out of the Bennett Dam, the velocities are low enough to allow for the formation of a stable surface ice cover. Although there may be some local reaches where some shoving may occur, the general freeze-up pattern is one where large sheets of surface ice (not rafts of ice flows from frazil production) juxtapose on the surface. Figure 21 illustrates the typical ice cover characteristics in this reach following freeze-up in 1993.

From a modelling perspective, this indicates that once the water temperature has fallen to zero, then ice forms and a stable cover can exist. Thus it is a simple matter to simply calculate the progression of the ice cover by calculating the time at which a water parcel moving through the system reaches zero degrees. Figure 22 illustrates these calculations and compares the modelled freeze-up date to the observed date of freeze-up at Peace Point.

5.2.2 Vermilion Chutes

The mechanism by which the ice cover progressed upstream of the Chutes is not yet well understood. Prior to this study, it was not known whether lodgement occurred independently upstream of the Chutes or if the ice staged over the Chutes once a stable ice cover had developed downstream of the Chutes. Observations carried out in the winter of 1993 focussed on identifying the mechanisms by which the ice cover progresses upstream over the Vermilion Chutes and measuring the ice characteristics in the vicinity of the Vermilion Chutes to determine the characteristics of both the ice cover and the river channel at that location.

The freeze-up process in the reach downstream of the Vermilion Chutes and the ice cover progression over the Chutes was observed from aircraft during the period between Nov 30 and Dec 12, 1993. In January, 1994 after the ice cover had stabilized, a field survey was undertaken to establish temporary benchmarks along the reach of the river between the mouth of the Mikkwa River and upstream of the Chutes. In addition, ice thickness measurements were made, the ice characteristics evaluated, and a water and bed level profile was established.

The Vermilion Chutes are composed of a set of rapids or falls located two kilometres apart (Figure 23). The upper discontinuity is described as being rapids, while the lower discontinuity is noted as a falls. Regardless, the Chutes provide an approximate drop of 9 m over a distance of about 3 kilometres. The rapids or the upper falls contribute to approximately 30% of the total drop, while the lower falls provide the other 70%. A significant scour hole has developed below each of the falls. Although not well documented, the upper scour hole is at least 5 m deep, while the scour hole below the lower falls has been measured to be about 20 m in depth (Figure 24). Obviously, there is sufficient turbulence at the base of each of the falls to prevent sediment deposition and to maintain the scour holes. On the other hand, the falls are turbulent enough to destroy the integrity of the incoming ice flows and cause hanging dams to develop in each of the scour holes (Figures 24 and 25).

The observations indicated that the ice cover upstream of the Chutes was explicitly linked to the formation of the ice cover downstream of the Chutes. The velocities at the base of the Chutes are too large to allow the juxtaposition of the ice cover to progress over the Chutes. Instead, the frazil pans which are destroyed in the rapids are transported under the cover and deposited along the underside of the stable ice cover downstream of the Chutes. The rate at which the ice is deposited depends on the supply of incoming ice and the rate at which it is transported along the underside of the ice cover. As long as the ice supply is greater than the transport rate under the ice the thickness of the ice accumulation will increase. When the additional thickness and roughness is sufficient to raise the water level downstream of the Chutes to a level equal approximately to the height of the Chutes, the ice cover can progress upstream by a variety of mechanisms, depending on the discharge and ice supply (Figure 26). Figure 27 illustrates an empirical relationship that defines the conditions which must be met before the ice cover can progress upstream over the Chutes. The air temperature reflects the ice supply and the discharge is an indication of the height which the ice cover below the Chutes must increased. Thus, as the air temperature decreases, for a given discharge, the time that it takes for the ice to stage over the falls decreases.

5.2.3 Vermilion Chutes to Taylor

5.2.3.1 Border Ice Growth Andres (1993) reported on measurements undertaken to verify a model used to calculate the growth rates of border ice. These appear to be the only records of border ice widths on the river. Unfortunately these measurements are limited to the area around and upstream of Peace River town. The measurements were made from an aircraft at four different times ove the course of freeze-up. Each of the measurement sites was located on 1:50,000 scale NTS maps and the channel width was estimated from the map. The width of the border ice was then scaled from the photograph and the percentage of border ice coverage (of the total width) was determined. The width of the border ice was extremely variable and seemed to be affected by local hydraulic conditions. Where the channel was well defined, straight, and with no islands to cause large eddy zones, the border ice growth was limited. However, in areas where the channel tends to be wider, such as in the vicinity of islands, or along large bends, the growth of the border ice was more substantial.

This suggests that the growth of the border ice tends to decrease the variability of the width of the water surface, when considered over a long reach. Furthermore, the border ice did not play a major role in the defining the ultimate characteristics of the ice cover. Even after three months of subzero temperatures, the width of the border ice was less than 15% of the total width.

5.2.3.2 Cover Formation Freeze-up on the Peace River between the Vermilion Chutes and Taylor occurs by a orderly progression of the ice cover. Water of temperature varying between 2°C and 9°C enters the river from upstream of the Bennett Dam. As the water moves downstream it cools, generates frazil and produces flowing ice, the concentration of which increases in a downstream direction. The rate at which the ice cover progresses upstream is related to the rate at which ice is generated and the thickness of the newly formed ice cover. The first factor depends on the discharge in the river, the rate of heat loss from the water surface and is related to the incoming solar radiation, air temperature, and wind. The second factor depends on the channel geometry, slope, discharge, and air temperature. Generally, the higher the discharge the thicker the accumulation of ice.

Alberta Environmental Protection (Fonstad and Garner, 1984, for example) has been monitoring the progression and recession of the ice cover since the mid 1970's. Figure 28 illustrates the extreme variability that the location of the head of the ice cover (hence the timing of freeze-up) can have in any given year over the period of the observations. Unfortunately, the observations do not include the type of ice cover, that is whether it is consolidated or juxtaposed and therefore the dominant freeze-up process only be inferred from the progression rates. That is, a large progression rate suggests that juxtaposition is the dominant mode of ice formation, whereas a low rate of upstream progression suggests that consolidation is the dominant process. This logic seems to work well downstream of Dunvegan. Upstream of that location, the proximity to the Bennett Dam and the variability in the flows and ice characteristics make it difficult to speculate about the dominant process.

This type of rationalization can be illustrated for the 1982 freeze-up (Figure 29). In 1982, the ice front was first observed at Fort Vermilion on November 23. With air temperatures in the order of

-15°C and discharges at Peace River fluctuating between 1600 and 2000 m^3/s, the ice front progressed upstream at an average rate of about 20 km/day, reaching Manning on December 9. The very mild slope of the river in these sub-reaches (Table 1) must have allowed a thin cover to form by juxtaposition, on some days. This resulted in a relatively rapid upstream progression of the ice cover.

At Manning, where the slope increases by about an order of magnitude (Table 1), juxtaposition was no longer possible and the cover formed only by shoving. At temperatures in the range of -20°C and discharges of about 1700 m^3/s, the rate of upstream progression slowed to about 4 km/day, with the ice front passing Peace River on January 4 and reaching Dunvegan on January 13. In the vicinity of Dunvegan, the rate of upstream progression increased again, suggesting that juxtaposition may have been responsible for the formation of the stable ice cover. This may have been brought about by lower than usual releases from the Bennett Dam. No observations were available after January 17, at which time the head of the cover was located about 30 km upstream of Dunvegan.

In 1993, work on another NRBS project provided an opportunity to characterize the type of ice cover that developed and thus explicitly identify the mode of ice cover development and determine the attendant stability criteria. An aerial reconnaissance was undertaken on February 10, 1993 in a reach of the river between the Shaftesbury Ferry and the mouth of the Notikewin River. The intention of the work was to infer, from the surface characteristics, the freeze-up mode and the relative differences in the potential thickness and roughness of the ice cover in each of the reaches. The surface of the ice was characterized as being either smooth, which is indicative of a juxtaposed ice cover, or rough, which describes an ice cover that has undergone substantial shoving or consolidation. Furthermore, the evidence of large rafts, which were imbedded within the ice cover was also noted, along with the existence of shear lines which are indicative of a consolidating ice cover. It should be noted that the shear lines may not have been related to the existing ice cover, but could have been relics of a previously formed cover that had collapsed prior to the formation of the observed ice cover. Regardless, their presence suggests that considerable thickening and storage of frazil has occurred and that there may be a reduction in the width or the cross sectional area due to the accumulated ice.

Figure 30 summarizes the discharge, air temperature, progression rate, and the surficial characteristics of the ice cover within the study area. The ice cover formed in the time period between Dec 21 and Dec 30, 1992. Downstream of the Whitemud River, the ice cover developed at discharges of about 1800 to 1900 m^3/s and at temperatures in the order of -25°C . This resulted in a cover that was generally rough due to shoving of the juxtaposed ice floes. Apparently the juxtaposed floes could not gain sufficient strength due to downward freezing at these temperatures before the stresses on the cover due to the high discharges and the lengthening ice cover increased to the point where the cover would collapse. Upstream of the Whitemud River, freeze-up occurred during much colder conditions (in the order of -40°C) and at discharges of about 1700 m^3/s. It appears that at the colder conditions and lower discharge, the cover gained sufficient strength from freezing that the juxtaposed floes were not consolidated or shoved as the pack progressed upstream.

As a result, the ice cover between the Shaftesbury Ferry and the Whitemud River was generally flat, composed of either individual pans or large rafts. The surficial roughness was largely due to the ridges which where produced when the individual pans impinged upon one another. The average thickness was about 1.1 m and the variability (defined by the standard deviation of the ice thickness measurements at each section) across the channel was about 0.22 m. Downstream of the Whitemud River to as far as the Notikewan River, the cover was generally rough. The surface roughness was due mostly to the underturning and shoving of the floes. The average thickness in this reach was about 1.5 m with a variability of about 0.55. It should be noted that the ice thickness measurements were undertaken about six weeks after freeze-up. This would allow for a substantial amount of redistribution of the frazil under the ice cover. That is, some of the areas where the ice was initially very thick and rough would thin out and become less irregular. On the other hand, the transport and deposition of frazil could cause some of the initially smooth areas to become thicker and more irregular.

The above observations confirm that discharge alone does not determine the characteristics of the freeze-up ice cover. Temperature also has a role in determining the strength of the cover as reflected by Equation [33]. A model developed by Andres (1988) accounts for these process and seems to be able to simulate the upstream progression of the ice cover for explicit boundary conditions and explicit daily values of air temperature, solar radiation, and discharge. Additional work carried out for the MBIS (Andres, in press) illustrated the effects of air temperature and discharge on the ultimate configuration and extent of the ice cover over the entire winter. The measured data shown in Figure 28 can be reproduced with relative accuracy, even considering the number of processes (lodgement at Peace Point, staging over the Vermilion Chutes, and juxtaposition and consolidation) that have to be explicitly modelled. Unfortunately, this method of analysis is too unwieldy for a simple determination of the dominant cover characteristics as a function of channel characteristics, temperature, and discharge. Hence the utility of Equation [33].

As was discussed in Section 4.2.4, the stability of a juxtaposed ice cover is largely a function of the strength of the frozen ice pack as reflected by its failure mode (crushing, bending, etc.). Values of the strength of the ice can be estimated from the literature, but the actual prescribed strength is a function of how the ice fails. Hence, there is a need to calibrate for the effective ice strength, if one failure mode is assumed in the stability analysis. Furthermore, the calibration process can identify the integrated effects of a number of other parameters such as the flow resistance of the ice cover, the lack of a perfect hydraulic model for flow under the ice cover, and errors in estimating the ice supply.

Post-regulation data from the three gauging sites (Tables 3, 4, and 5) and from other ice surveys for which the relevant parameters could be determined (Figure 30) were used in the calibration. Figure 31 illustrates a plot of the stability parameter $(Q/W)^{8/5}$ (L_c/T_a) $S^{1.7}$ as a function of the stage increase at freeze-up. Each data point was characterized as either being a consolidated cover, a juxtaposed cover, or a transitional cover on the basis of either direct observation or an interpretation from the estimated thickness. The very thick covers were assumed to be consolidated, the thin covers were juxtaposed, and those for which the cover characteristics were not obvious were termed transitional. Table 8 summarizes the data in Figure 31 and identifies the probability of either a juxtaposed or consolidated ice cover forming for a given stability parameter.

Table 8 **Probability of a given type of ice cover forming for a given stability parameter**

Stability Parameter	Probability of Juxtaposition (%)	Probability of Consolidation (%)
< 0.0005	100	0
0.0005 - 0.001	86	14
0.001 - 0.003	50	50
0.003 - 0.006	23	77
0.006 - 0.008	25	75
> 0.008	0	100

From the plot and Table 8 it is evident that critical value of the stability parameter is about 0.003. Below this number, the ice cover is predominantly thin, suggesting juxtaposition, and above this number the ice cover is predominantly thick, suggesting consolidation. Exceptions do arise, but these are probably attributed to an inaccurate estimate of the ice supply. For typical values of floe porosity and Manning's roughness the effective strength of the solid ice contributing to the frozen pack is about 700 kPa. This low value probably reflects the fact that the ice cover does not fail by crushing but by some other mechanism that does not allow for the full development of the crushing strength.

5.2.2.3 Ice Thickness If the cover forms by juxtaposition, its thickness should be close to the thickness of the pans that arrive at the head of the cover. Additional thickening can occur without consolidation due to the redistribution of the frazil. Very little is known about this process, except to say that the source of the redistributed frazil is in the high velocity areas, and the depositional zones tend to be low velocity areas such as backwater zones in the lea of islands or along the bank.

More is known about the characteristics of a consolidated ice cover. Work at the Peace River town by Neill and Andres (1984) and Andres (in press) has allowed for the characterization of the roughness and the internal strength of the consolidated ice cover. The roughness of the underside of the ice was found to be similar to the bed roughness and it was argued that it should be independent of the thickness because the characteristics of the interface is only determined by the characteristics of the frazil and the flow condition at the ice/water interface. That is, the ice roughness should be the same for both a juxtaposed ice cover and a consolidated ice cover. The work also showed that the coefficient of internal friction of a juxtaposed ice cover varied between 0.8 and 2.0 and averaged between 1.0 and 1.5. The lower value is more appropriate because the higher values are probably in part due to freezing in the pack. Figure 32 illustrates the range in the calculated dimensionless coefficient. Finally, the calculations suggested that the ice thickness in a consolidated cover was a function of the discharge, it varied between 1 and 4 m in thickness, and that the thickness estimates were in the same range as WSC measurements later in the winter.

6.0 EFFECTS OF REGULATION

6.1 FREEZE-UP DATES AND ICE THICKNESS

As mentioned earlier, the effect of regulation on the date of freeze-up and hence the duration of the ice covered conditions is to generally delay the freeze-up date and reduce the length of time over which an ice cover is in place. This effect tends to be less noticable the further downstream one gets from the Dam. For example, at Peace Point, which is representative of Reach 1 there has been at most a one week delay in the time a stable ice cover forms. In fact, at this location, although the pre- and post-regulation freeze-up dates are different, it would be difficult to prove conclusively that this shift was dependent on more than just climatic effects. Regardless, after regulation, freeze-up in Reach 1 usually occurs during the month of November.

At Fort Vermilion, the date at which an ice cover forms is a function of the time at which the ice cover can stage over the Vermilion Chutes. This depends on the supply of ice arriving at the Chutes and the discharge. No data could be found on pre-regulation freeze-up dates, however if the model developed at Peace River town is used, one would expect that freeze-up in the reach probably occurred around the same time as at Peace River. After regulation, the data suggest that freeze-up can occur any time between the middle of November and late December. This same variability in the freeze-up date applies to the entire river upstream of Fort Vermilion, except the mean freeze-up date gets later in the winter as one gets closer to the dam.

At the upstream end of Reach 2 freeze-up usually occurs between late November and early January. In Reach 3 freeze-up occurs between middle December and early January. This is typically a month or two later than what was occuring prior to regulation. At Peace River (located in the middle of Reach 4) the date of freeze-up has been set back by anytime between one week and two months from the normal pre-regulated freeze-up date of late November or early December. At Dunvegan, the upstream end of Reach 4, freeze-up does not usually occur until early January, and on one or two occasions a stable ice cover has not formed at all.

The biggest impact on the time of freeze-up generally occurs upstream of the BC-Alberta border. For about 60% of the years, the ice cover has not progressed up to this location. When a stable ice cover does form, it only exists for about one month, or about 20% of the time prior to regulation. Table 9 summarizes this data for some salient locations along the river.

The ice thickness at any particular location is a function of the discharge and air temperature at the time of freeze-up. Thus it is difficult to prescribe a typical ice cover on a reach by reach basis. Even for a constant discharge, the daily variation in the air temperature will result in a different form of the ice cover, as defined by the stability equation, at different locations within each reach. Figure 33 illustrates the effect of air temperature and discharge on the cover characteristics for each of the reaches. It is evident that a juxtaposed ice cover should form in Reach 1 and 2 (downstream of the Notikewin River) under all discharges that one could expect from the Bennett Dam, regardless of the air temperature. This would produce an ice cover with a thickness of about 0.50 m (the thickness of the frazil floes) unless there happened to be a dramatic loss in stability because of a surge of ice and water from upstream.

Table 9 **Summary of post-regulation freeze-up dates and duration of the ice cover**

Location	Average Freeze-up Date	Average Duration of the Ice Cover (months)
Fort Vermilion	Dec 1	5
Notikewan River	Dec 15	4
Peace River	Jan 1	4.5
Dunvegan	Jan 15	3
BC/Alberta Border	Feb 15	1
Taylor	no ice	0

Upstream, in Reach 3 and 4, it is evident that a variety of ice covers could form for typical releases from the Bennett Dam, depending on the air temperature. For example, at a discharge of 2000 m^3/s a juxtaposed ice cover could form only when the air temperature was near -30°C. At temperatures between -10 and -20°C, any discharge above 1200 to 1500 m^3/s would result in a consolidated ice cover. Under typical winter temperatures, the discharge must be less than about 800 to 1000 m^3/s to ensure a juxtaposed ice cover. In Reach 5 and 6, the slope is so steep that a juxtaposed ice cover will not form even on the coldest days unless the discharge is below 1000 m^3/s.

Figures 34 and 35 illustrate the modelled ice thickness and the stage increase that are associated with a consolidated ice cover in Reach 3, 4, 5, and 6. Both are a function of discharge, the ultimate thickness reflects strongly the slope of the reach, and the stage increase is similar to that measured at the WSC gauges. It should be noted that the thickness was calculated by assuming that the value of the dimensionless coefficient of internal friction is 0.8. Depending on the discharge, the stage increase varies from 3.5 to 5 m in Reach 3, 4 and 5 and from 4.5 to 6 m in Reach 6 if an ice cover should happen to form in that reach. The ice thickness exhibits similar trends.

6.2 EFFECTS ON DISCHARGES

As mentioned earlier, the winter discharges in the river are largely controlled by releases from the dam. These releases typically vary between 1500 and 2000 m^3/s, and because of the continuity and limited inflows from tributaries, there is not much variation along the entire river. However, considerable flow can be lost to storage as the ice cover forms and progresses downstream. During juxtaposition, the ice cover can abvance by up to 50 km/day. This results in a loss to channel storage of up to about 30% of the discharge arriving at the ice front. On the other hand, when a long length of cover consolidates, an equivalent amount of flow can be released from storage. This flow increase may be sufficient to destroy the ice cover downstream and cause a dramatic increase in both ice thickness and water levels downstream. At least two such events have been experienced at Peace River town in 1982 and 1992, and it would be expected that this phenomenon probably occurs each year at some location along the river.

A result of the effort by the ice cover to achieve some sort of equilibrium is that water levels, ice thicknesses, and discharges can change dramatically at any location along the river within 100 km of the ice front while the ice cover is in its formative stage. This makes it difficult even to estimate flows along the river during this period as well as determining minumum and maximum design flows for any number of physical and biological processes.

6.3 IMPACTS ON HABITAT AND GROUNDWATER LEVELS

It is beyond the scope of this study to evaluate the impacts of regulation on either habitat or flood levels, other than to illustrate the impacts the changed ice regime may have on the channel itself. These effects are most apparent upstream of the Vermilion Chutes. Certainly, the extent of the open water season has been changed dramatically in the reaches upstream of the Notikewin River. This will substantially decrease the opportunuity for movement across the river, especially upstream of the BC/Alberta border where it effectively has been removed.

The very thick ice cover that forms in these upstream reaches potentially has an effect on the fish habitat in the shallow water zones around islands and near the bank. One of the effects of the deposition of large amounts of frazil is to narrow up the channel over the entire winter. The thick frazil tends to stay in place in the backwater areas and along the bank while the frazil thins out in the higher velocity areas. This becomes somewhat of a non linear process, and as the high velocity zones thin out, more flow is concentrated in those zones and less flow is evident in the shallower areas. Thus a "conduit" of relatively high flow exists in the centre of the channel, flanked by significant areas full of frazil with little or no flow. If one looks at a map of the channel planform, it will be evident which areas will exhibit large accumulations of frazil and where the main flow area will be.

High water levels are generally associated with a thick ice cover. Again, in the reach upstream of the Notikewin River, to about Dunvegan, thick ice covers that produce high water levels can exist for relatively long periods of time (two to three months) compared to the duration of an open water flood with the same stage increase. This can result in a supercharging of the groundwater conditions along the margins of the river. A very visible example is West Peace in the town of Peace River (G. Fonstad, personal communication). At that location the high groundwater levels lead to the flooding of the basements of some of the residences. This can persists for much of the winter. It should be noted that the water loss from the river due to this phenomenon is very small during the surcharged condition and the effects of the high groundwater appear to dissipate after the ice is removed during breakup. There does not appear to be any apparent cummulative effects on the long term groundwater levels from the winter surcharging.

7.0 SUMMARY AND RECOMMENDATIONS

This report has reviewed the processes by which an ice cover forms on large regulated and non-regulated rivers. Explicit equations and algorithms have been presented that quantify these processes. Work that had been undertaken previously on the Peace River was also described to provide a framework for the calibration of these algorithms for the Peace River in its regulated and non-regulated condition. The significant theoretical advances that were made include the development of a procedure to forecast freeze-up on a non-regulated river and the derivation of a stability relationship that uses both air temperature and discharge to determine whether a juxtaposed or consolidated ice cover will form. The latter development is important to characterize the type of ice cover that will occur on the Peace River under regulated conditions.

In addition to the work on the ice process, the hydraulic characteristics of the Peace River were evaluated for six distinct reaches between the Slave River and Taylor using the existing data base. The climatological characteristics of the basin were summarized, along with a description of the spatial and temporal variation in the flows for the periods before and after regulation.

The main results of the study are as follows:

1. Discharges following regulation are, on the average, about two to three times greater than those prior to regulation.

2. Prior to regulation, the river cooled from a maximum annual water temperature of about 22°C to 0°C at the same rate as the declining air temperature. Ice began to form in early November, in most years, and an ice cover formed due to lodgement when the discharge deceased sufficiently to reduce the width of the flow by about 10%. A stable ice cover usually formed in early November at Peace Point and in late November or early December at Peace River. There is no data for Taylor, although the freeze-up probably occurred in early December.

3. After regulation, the high discharge of relatively warm water from upstream of the Bennett Dam has delayed the time of freeze-up and shortened the duration of the ice cover significantly in the reaches upstream of Fort Vermilion. At Taylor, and upstream of the BC/Alberta border, an ice cover is an exception rather than a rule. At Peace River, and downstream to Fort Vermilion, the freeze-up date has been delayed by as much as one to two months. Regulation has appeared to have only minor effects on the date of freeze-up downstream of the Vermilion Chutes and at Peace Point.

4. The ice cover downstream of the Notikewin River (Reach 1 and 2) generally forms by juxtaposition even after regulation due to the mild slope of the river. The ice cover thickness in these two reaches is only about 0.5 m thick, immediately after freeze-up and the stage increase associated with freeze-up is only about 1 to 2 m. The increase in stage is due mostly to the additional flow resistance of the ice cover. In the reaches between the Notikewin River and Dunvegan, where higher slopes are evident, either a juxtaposed or consolidated ice cover can form, depending on the discharge and the air temperature. For typical post-regulation discharges, the air temperature must be at least -30°C for a juxtaposed cover to form. To ensure that a juxtaposed

ice cover forms, for the typical range of temperatures expected during the winter, the discharge should be less than 800 to 1000 m^3/s. The stage increase under a juxtaposed ice cover is less than 2 m, while for a consolidated ice cover the stage increase can be as great as 5 m, with an ice thickness of about 4 m.

5. Between Hudson Hope and Dunvegan, the steeper river slopes prevent the formation of a juxtaposed ice cover under any combination of discharge or air temperature. Although the development of an ice cover in these two reaches is infrequent and when it does occur its duration is short lived, the formation thickness can approach 5 m and the increase in the stage can be up to 6 m.

6. The main physical impacts on the environment that have been identified relate primarily to (1) the existence of high water levels for long periods of time in areas where a consolidated ice cover has developed, (2) the losses in up to 30% of the flow into channel storage as the ice cover advances, (3) the potential unstable water levels and ice thicknesses that are evident within 100 km of the advancing ice cover, (4) the reduction in the duration of an ice cover for most of the length of the Peace River, and (5) dramatically thicker deposits of frazil in low velocity areas of the river upstream of the Vermilion Chutes. The high water levels associated with a consolidated ice cover produce both high groundwater levels and river levels which can persist for long periods of time but seem not to produce any long term cumulative effects. It should be noted that water levels related to the consolidated ice covers are in the same order as the maximum levels associated with the largest open water floods and, although they occur much more frequently, they are localized. The losses to channel storage, which can be as high as 500 m^3/s can effect the calculations of minimum flows necessary for a variety of physical and biological processes. The lack of an ice cover over 100-200 km lengths of the river will reduce the natural access across the river that was available prior to regulation. The thick deposits of frazil will reduce or eliminate flow in many of the shallow areas around the islands and near the edge of the bank.

Although algorithms have been developed for many of the processes identified on the Peace River, additional work is required to improve the modelling capabilities. It should be noted that this modelling framework will be an important tool to optimize the management of the river during winter conditions, and therefore it is important that the models can be made to operate as rigorously as possible. It is recommended that the following work be undertaken:

1. Additional observations and measurements be carried out downstream of the Vermilion Chutes to characterize better the freeze-up process in that reach. Furthermore, it is recommended that the natural flow, freeze-up model be calibrated and verified on the Athabasca River, the last remaining large river in Alberta that is well gauged and that has natural flows.

2 A model be developed to simulate the progression of the ice cover over the Vermilion Chutes. This process must be better quantified to improve the boundary and initial conditions for modelling the ice progression between Fort Vermilion and Taylor. It is also suggested that the bench marks established around the Vermilion Chutes, as part this study, be referenced to a common datum. This will improve the understanding of the hydraulics of the Chutes.

3. More work is required to verify the stability criteria used in determining the dominant mode of cover formation. An important component of this work will be the unsteady simulation of a consolidation event, and ultimately the development of an unsteady model to simulate the upstream and downstream progression of the ice cover.

4. Some effort must be expended to explicitly model the formation of frazil ice flows. This is important to determine their ultimate thickness and porosity and is crucial to modelling the ice discharge, the thickness of a juxtaposed ice cover, and the redistribution of frazil under the cover.

8.0 REFERENCES

Anderson, E.R. (1954) Energy budget studies in water loss investigations: Lake Hefner Studies. U.S. Geological Survey Professional Paper 269, Washington, D.C.

Andres, D.D. (1984) Thermal breakup predictions on a regulated river. Proceedings of the Conference on Water for Resource Development, ASCE, Coeur d'Alene, Idaho, pp. 534-538.

Andres, D.D. (1988) Observation and prediction of the 1986 breakup on the Athabasca River upstream of Fort McMurray, Alberta. Alberta Research Council Report Number SWE 88- 03, Edmonton, Alberta

Andres, D.D. and Spitzer, M.O.(1989) Inter-station storage losses during freeze-up on large rivers. Proceedings of the Seminar on Ice Jams and Flooding International Association for Hydraulic Research, XXIII Congress, Ottawa, Ontario.

Andres, D. D. (1993) Effects of climate change on the freeze-up regime of the Peace River: Phase I, Ice production algorithm development and calibration. A Report for the Mackenzie Basin Impact Study, Canadian Climate Centre, Environment Canada. Report No. SWE 93/01, Alberta Research Council, Edmonton, Alberta

Andres, D. D. (in press) Effects of climate change on the freeze-up regime of the Peace River, Final Report. A Report for the Mackenzie Basin Impact Study, Canadian Climate Centre, Environment Canada. Trillium Engineering and Hydrographics Inc., Edmonton, Alberta

Andres, D. D. (in press) A Case Study: Freeze-up and jamming on the Peace River at Peace River. In "The Monograph on River Ice Jams". S. Beltaos, Editor, Water Resources Press, Fort Collins, Colorado

Ashton, G.D. (1983) Lake ice decay. Cold Regions Science and Technology, Volume 8, pp. 83-86.

Bolsenga, S.J. (1964) Daily sums of global radiation for cloudless skies. U.S. Army Cold Regions Research and Engineering Laboratory Report 160, Hanover, New Hampshire.

Gray, D.M. (1970) Handbook on the principles of hydrology. National Research Council of Canada.

Hausser, R., J.P. Saucet, and F.E. Parkinson (1984) Coverage coefficient for calculating ice volume generated. Proceedings of the Workshop on Hydraulics of River Ice. Fredericton, New Brunswick, pp. 211-224.

Fonstad, G. and L. Garner (1984) Summary report, Peace River ice observations. Technical Services Division Report. Alberta Environment, Edmonton, Alberta.

Kellerhals, R., C.R. Neill, and D. Bray (1972) Hydraulic and geomorphic characteristics of rivers in Alberta. River Engineering and Surface Hydrology Report 72-1, Alberta Research Council, Edmonton, Alberta.

List, R.J. (1963) Smithsonian meteorologic tables. Smithsonian Institute, Washington, D.C.

Matousek, V.(1984) Regularity of the freezing-up of the water surface and heat exchange between water body and water surface. IAHR International Symposium on Ice, Hamburg, West Germany.

Neill, C.R. and Andres, D.D. (1984) Freeze-up flood stages associated with fluctuating reservoir releases Proceedings of the Cold Regions Engineering Specialty Conference, April 4-6, 1984. Canadian Society for Civil Engineering, Montreal, Canada.

Newbury, R.W. (1968) The Nelson River: A study of subarctic river processes. Ph.D. thesis, John Hopkins University, Baltimore, Maryland.

Pariset, E., R. Hausser, and A. Gagnon (1966) Formation of ice covers and ice jams in rivers. Journal of the Hydraulics Division, ASCE., Vol. 92, No. HY6, pp. 1-24.

Rivard, G., B. Michel, and F. Fonseca (1986) Formation and growth criteria for different types of ice in the St. Lawrence River. Proceedings of the Workshop on Hydraulics of River Ice, Montreal, Quebec, pp. H2.1-H2.10.

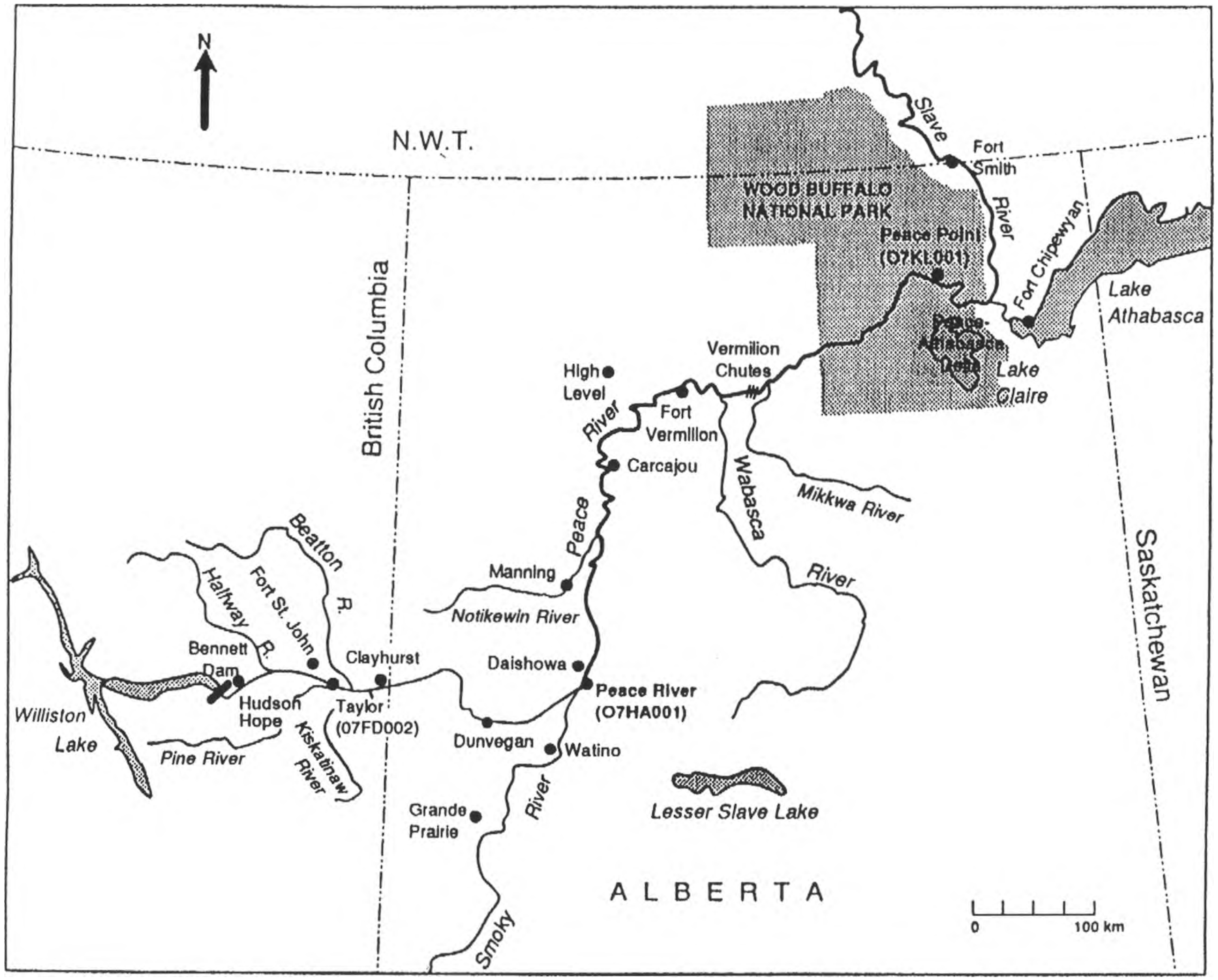


Figure 1 Study area

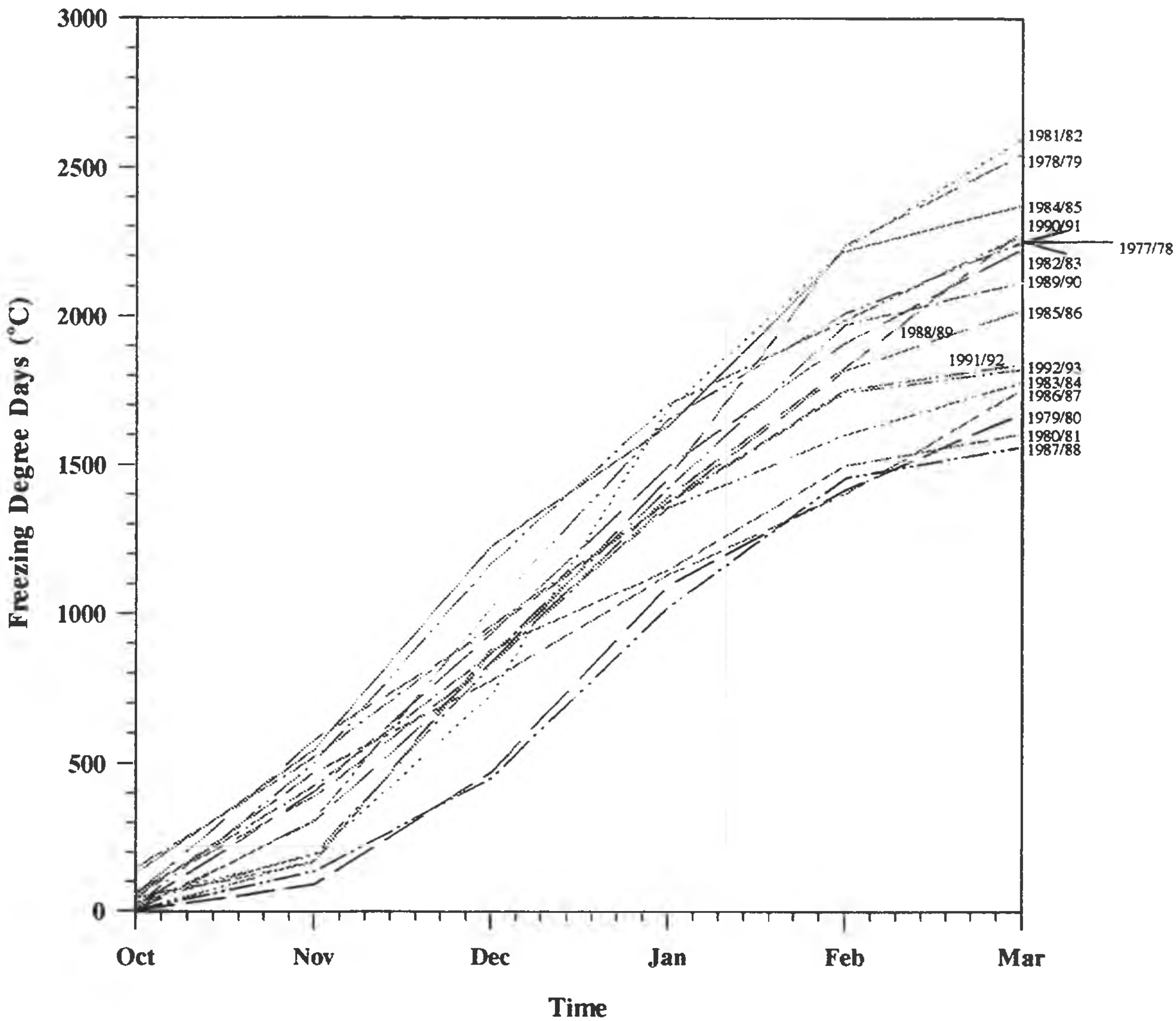


Figure 2 Variation in the accumulated degree days of freezing for the basin (1977 to 1994)

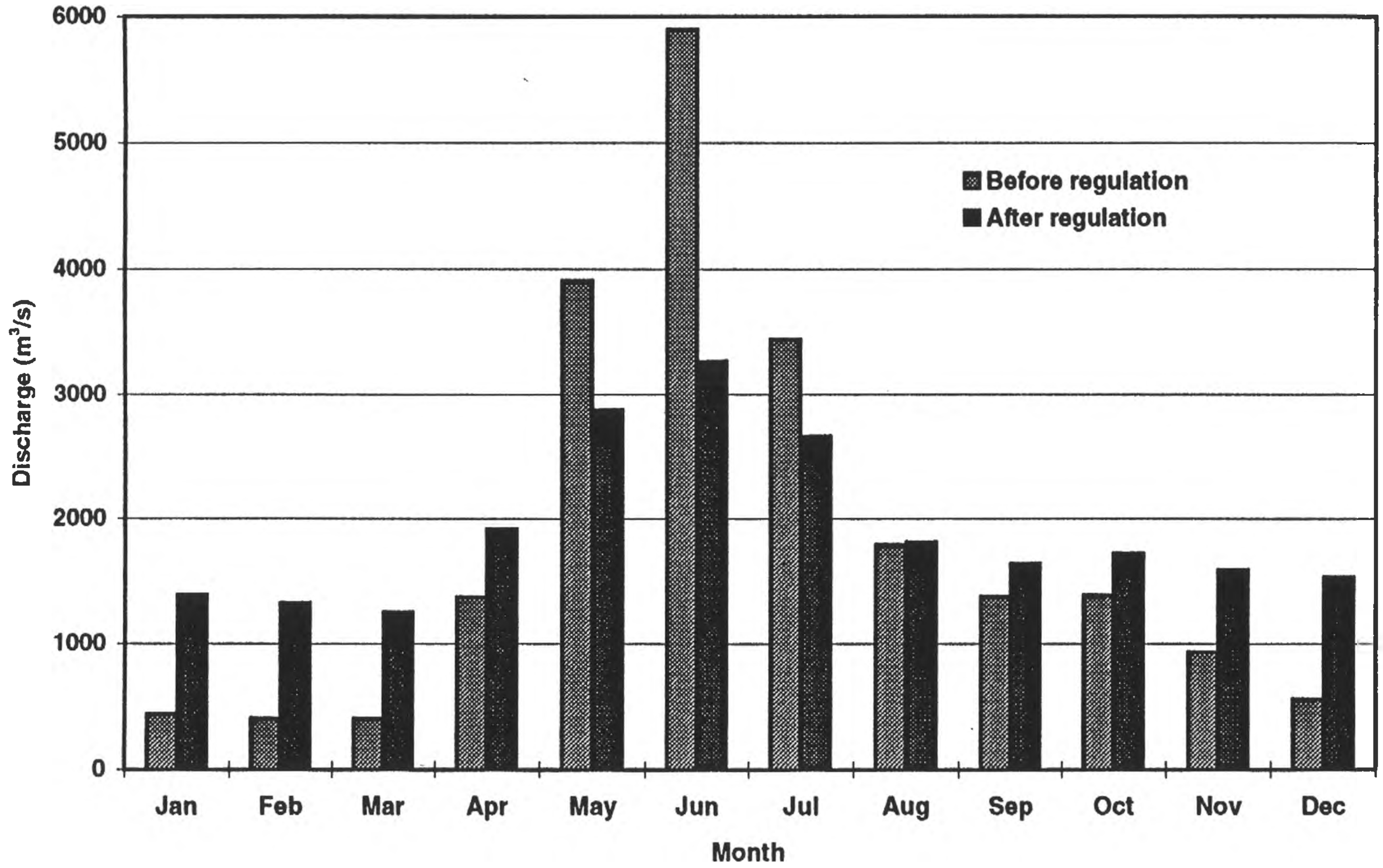


Figure 3 Comparison of pre- and post-regulation discharges at Peace River

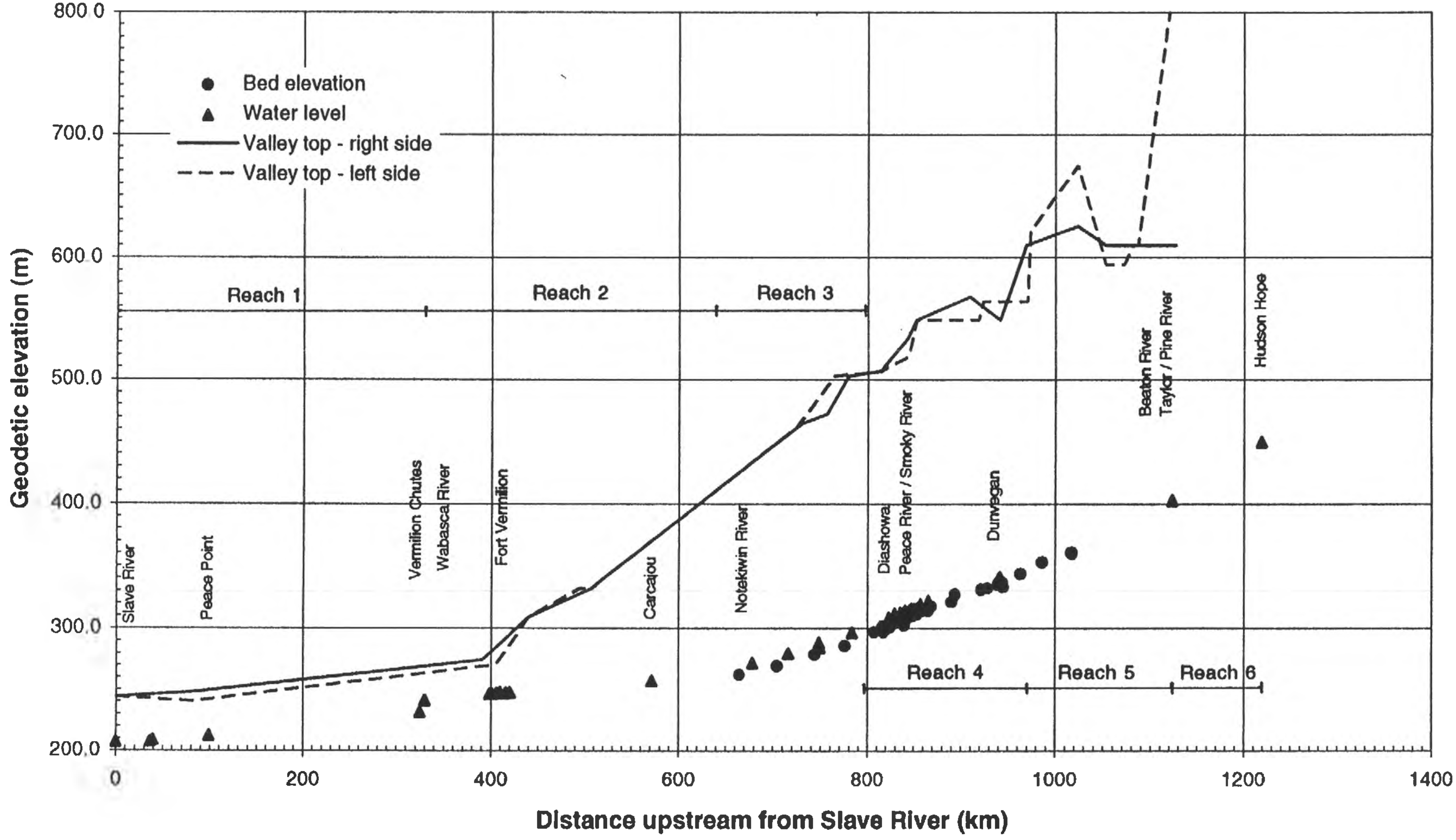


Figure 4 Profile of the Peace River, Taylor to the mouth.

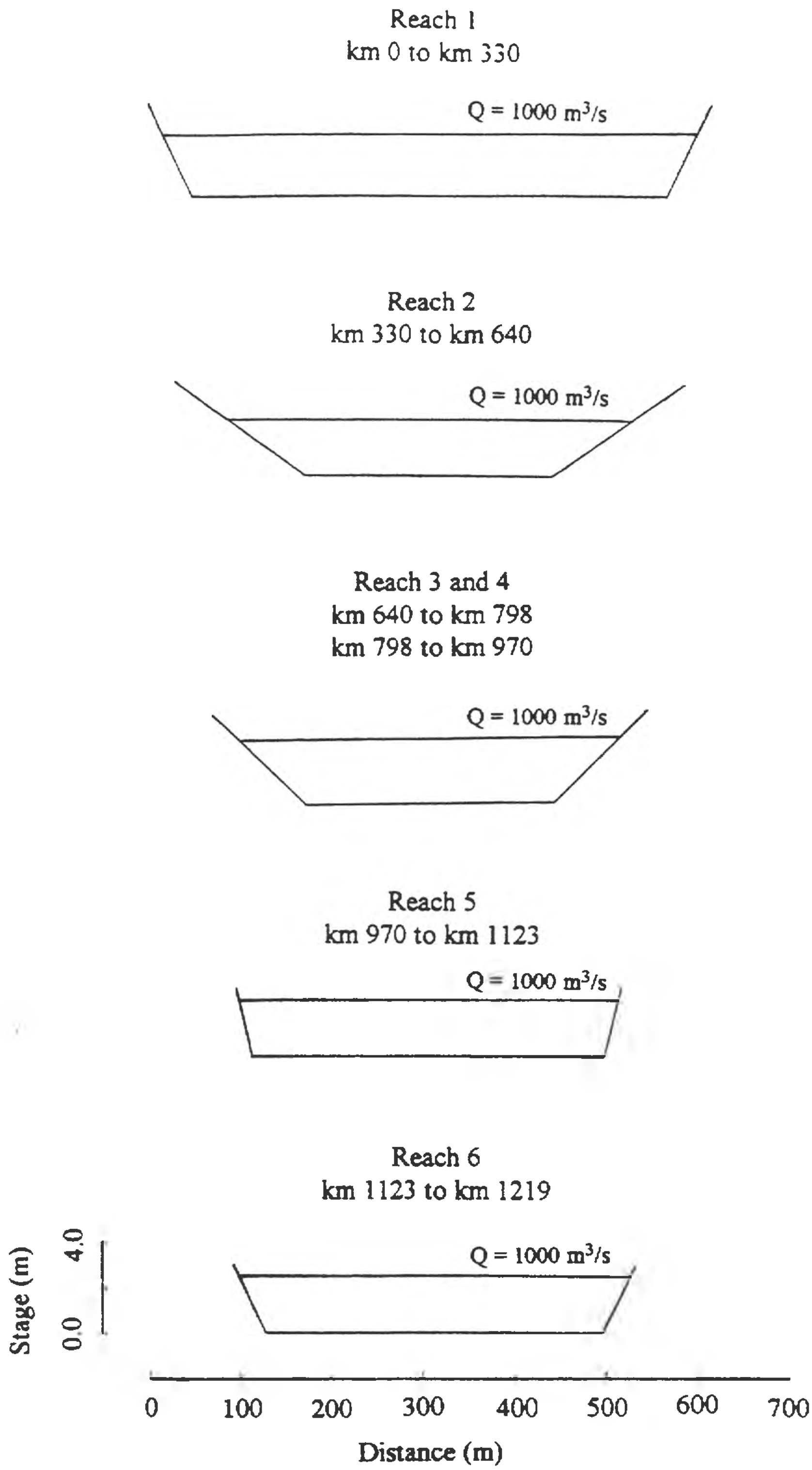


Figure 5 Adopted channel cross sections

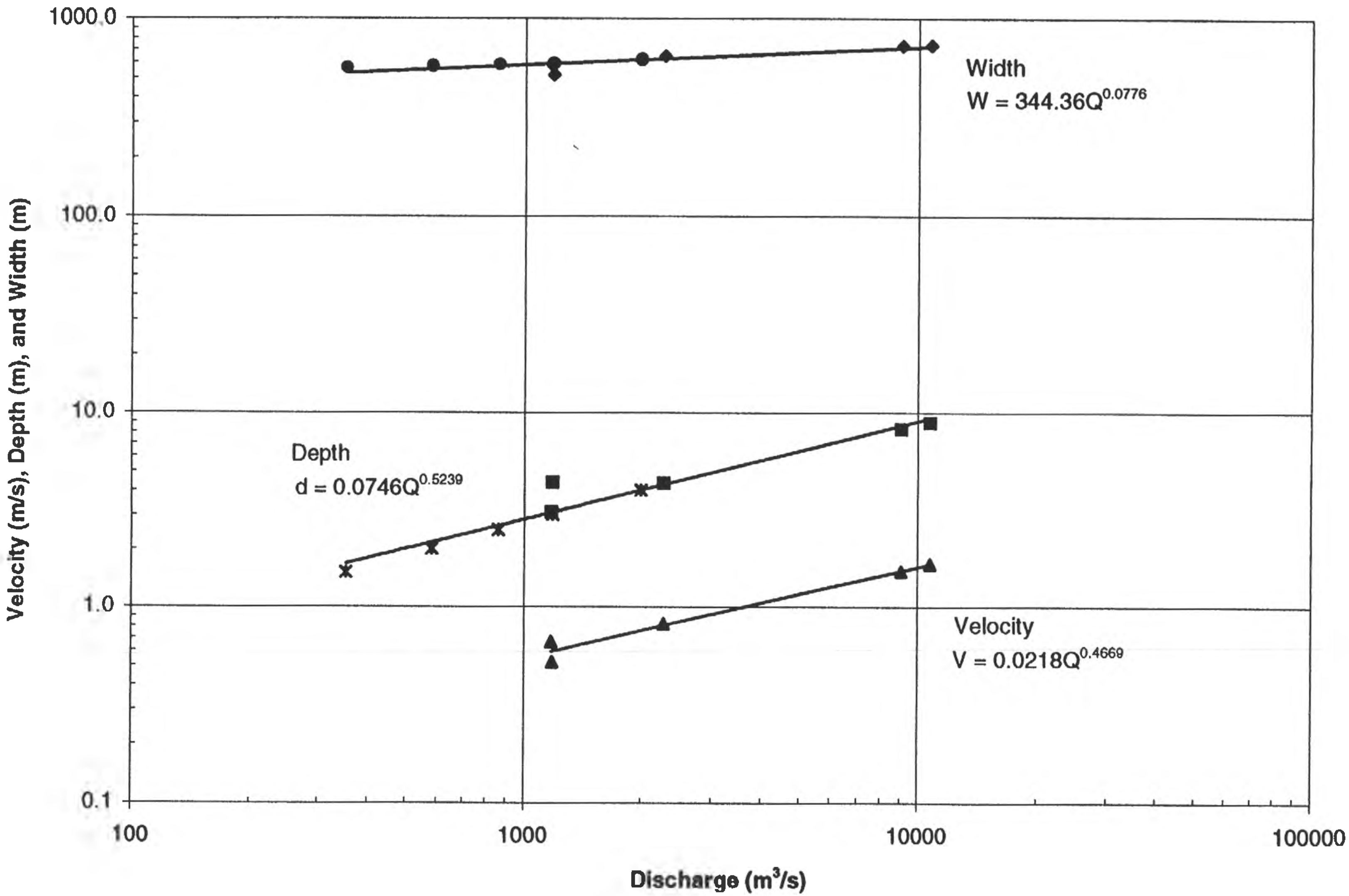


Figure 6 Hydraulic characteristics of reach 1

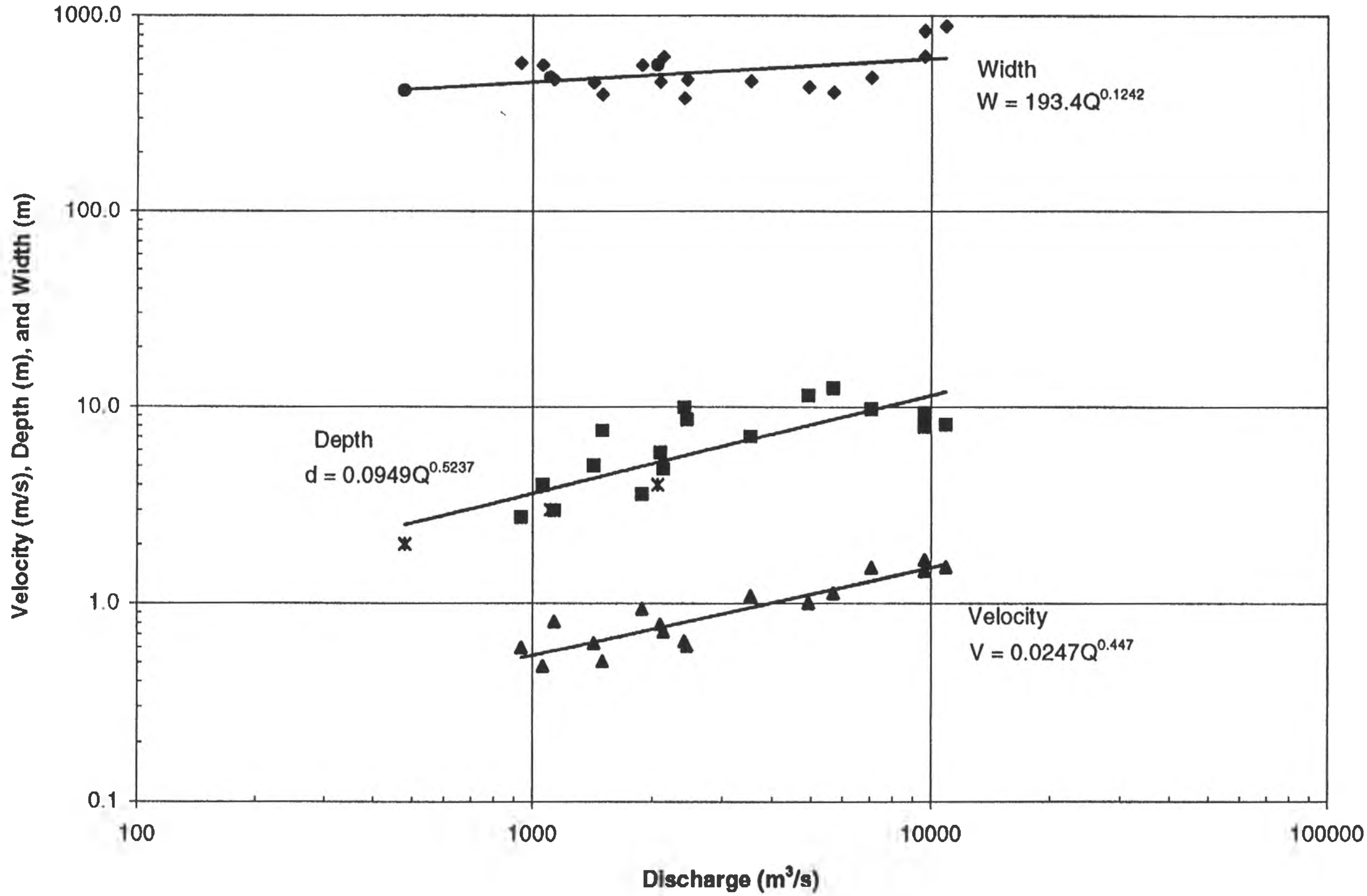


Figure 7 Hydraulic characteristics of reach 2

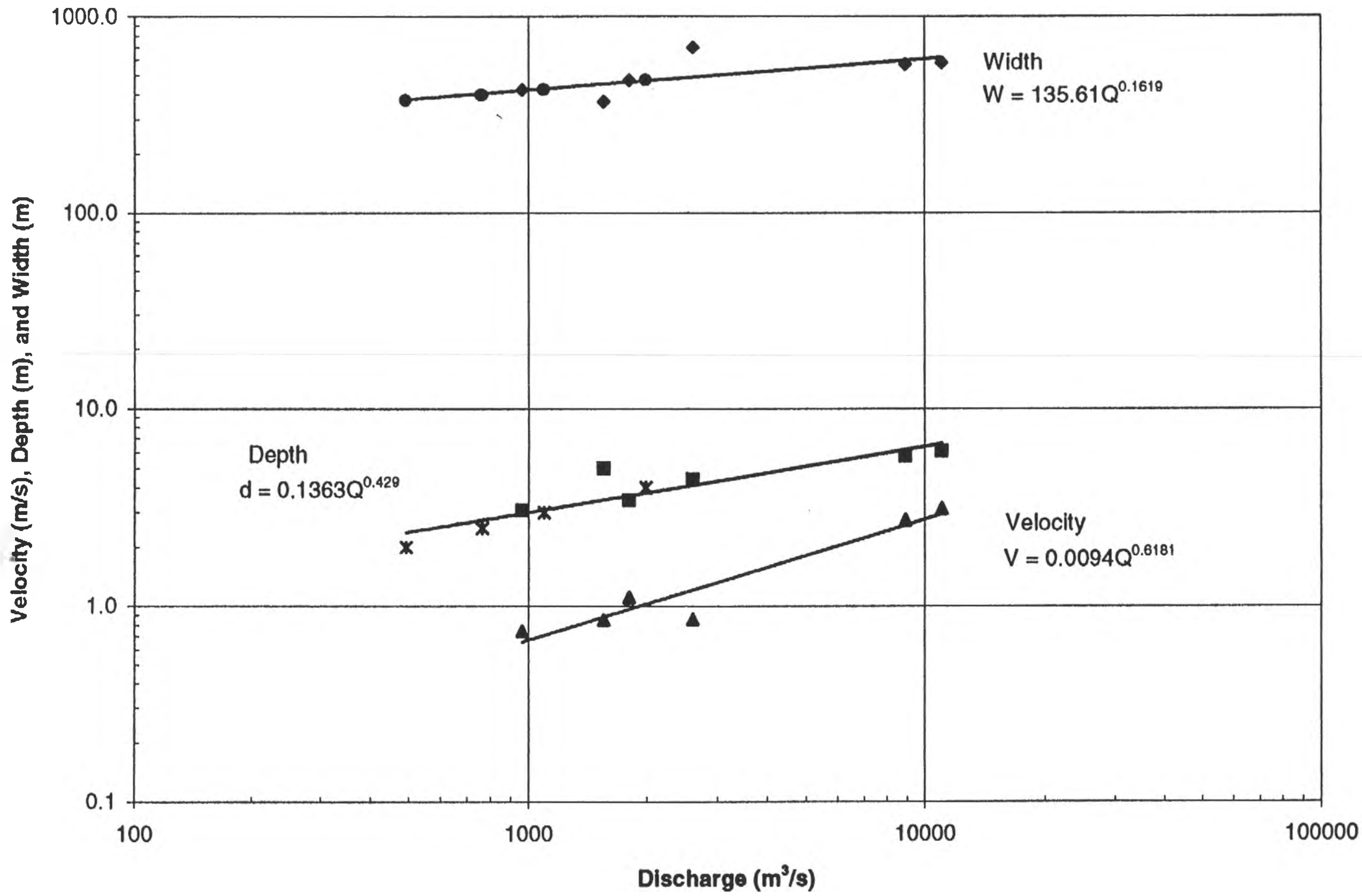


Figure 8 Hydraulic characteristics of reaches 3 and 4

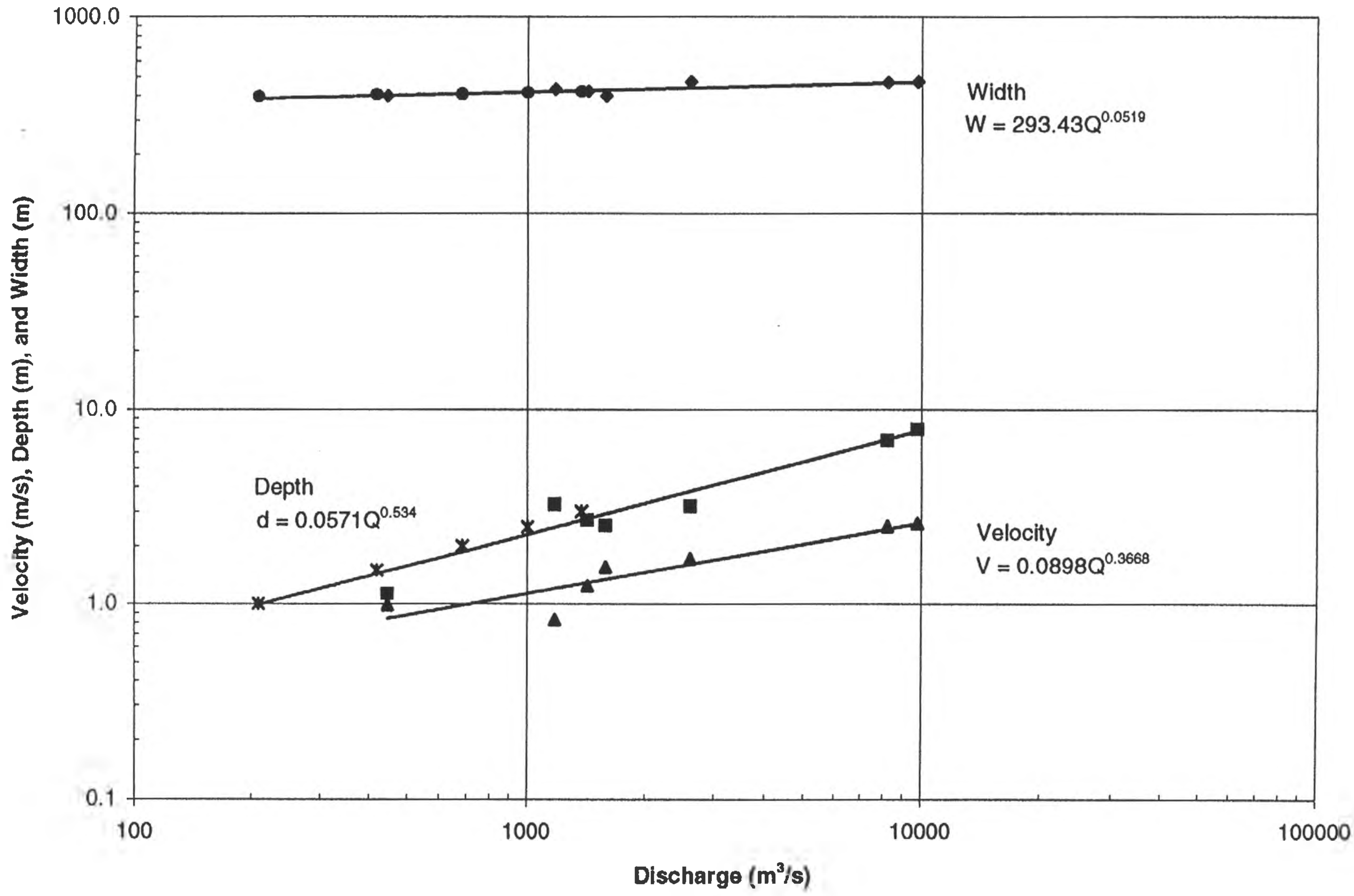


Figure 9 Hydraulic characteristics of reach 5

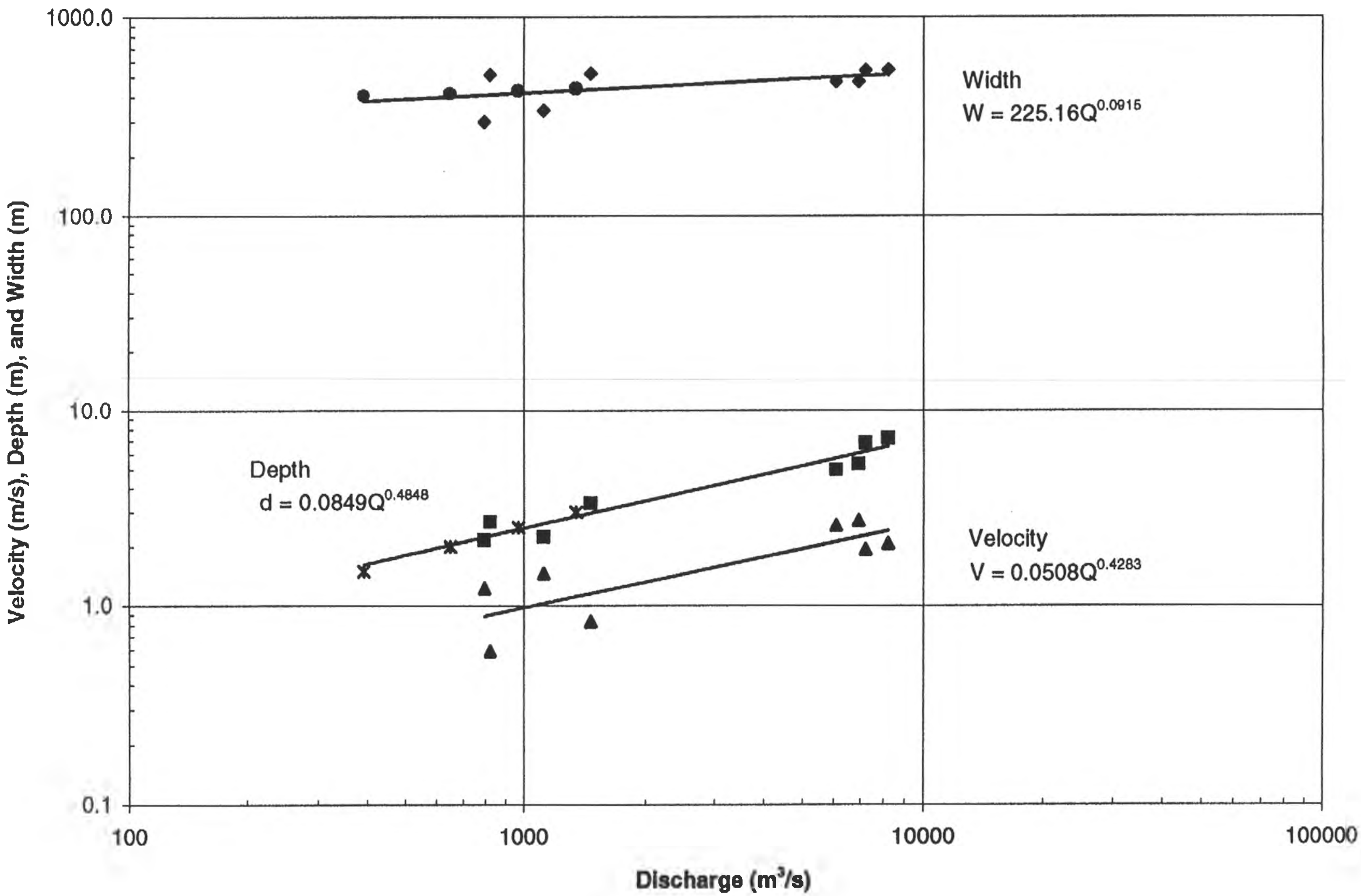


Figure 10 Hydraulic characteristics of reach 6

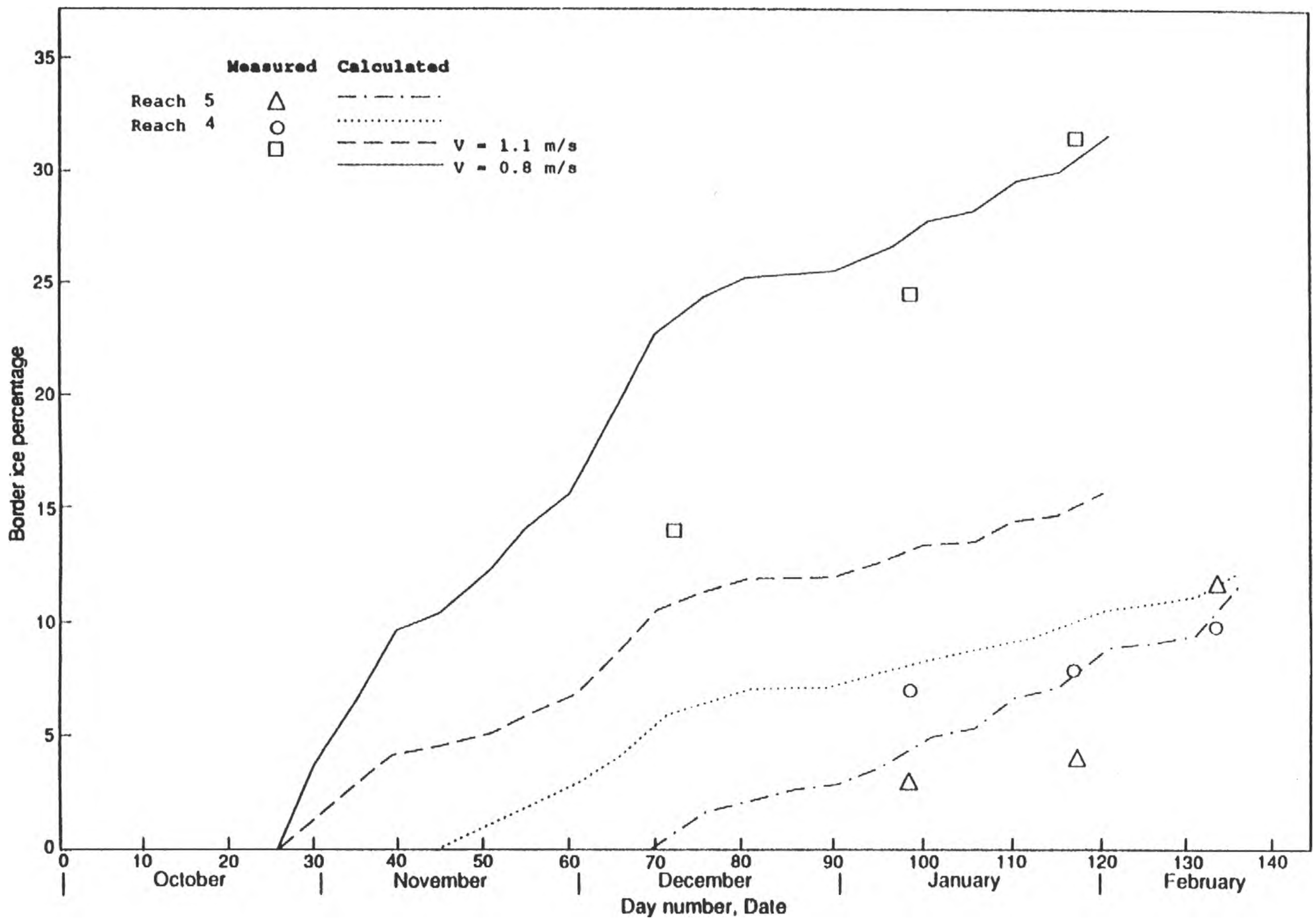


Figure 11 Border ice growth rates on large rivers

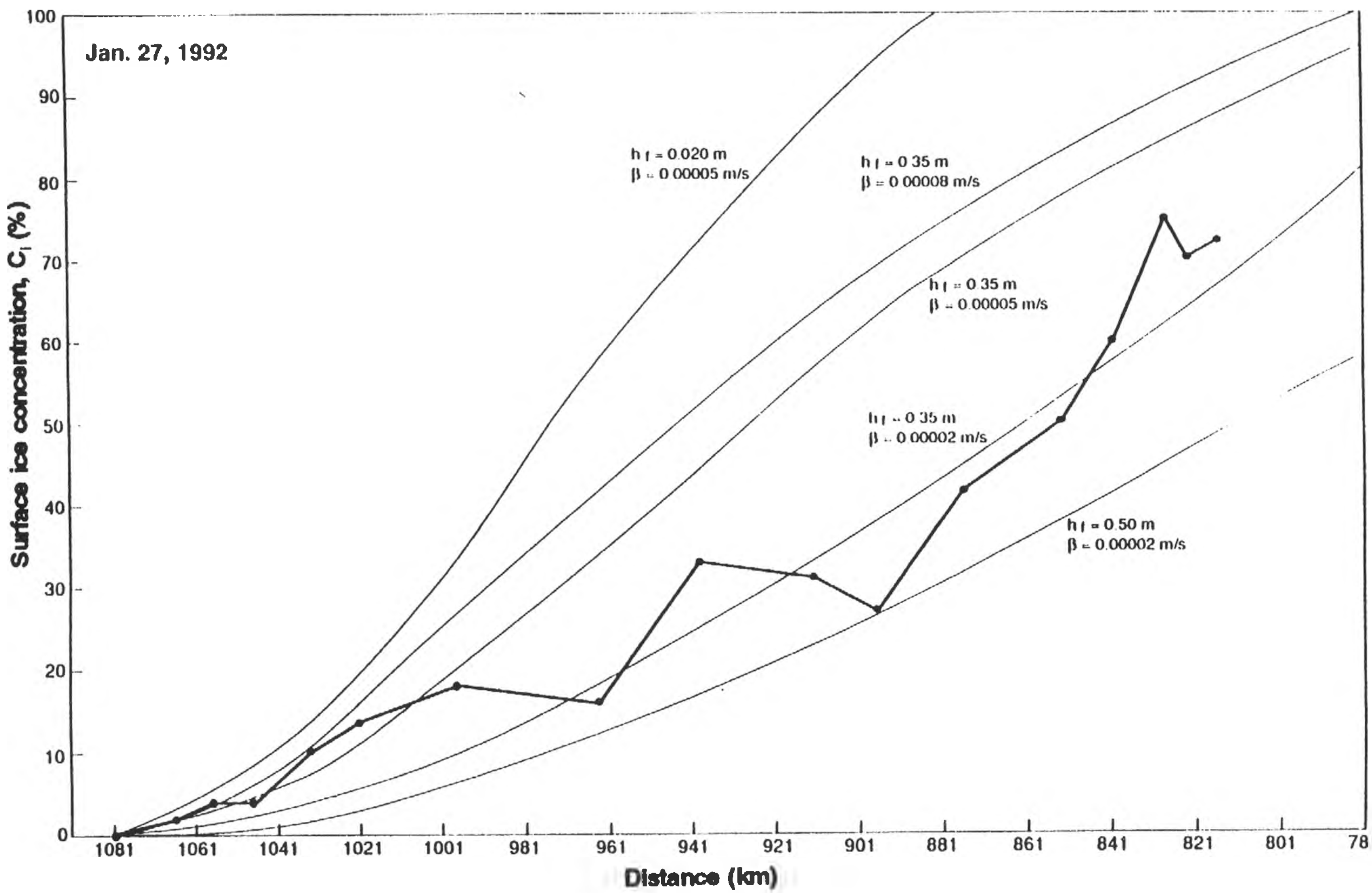


Figure 12 Sensitivity of surface ice concentration to changes in the floe thickness and rise velocity

Figure 13 Typical ice run during freeze-up.

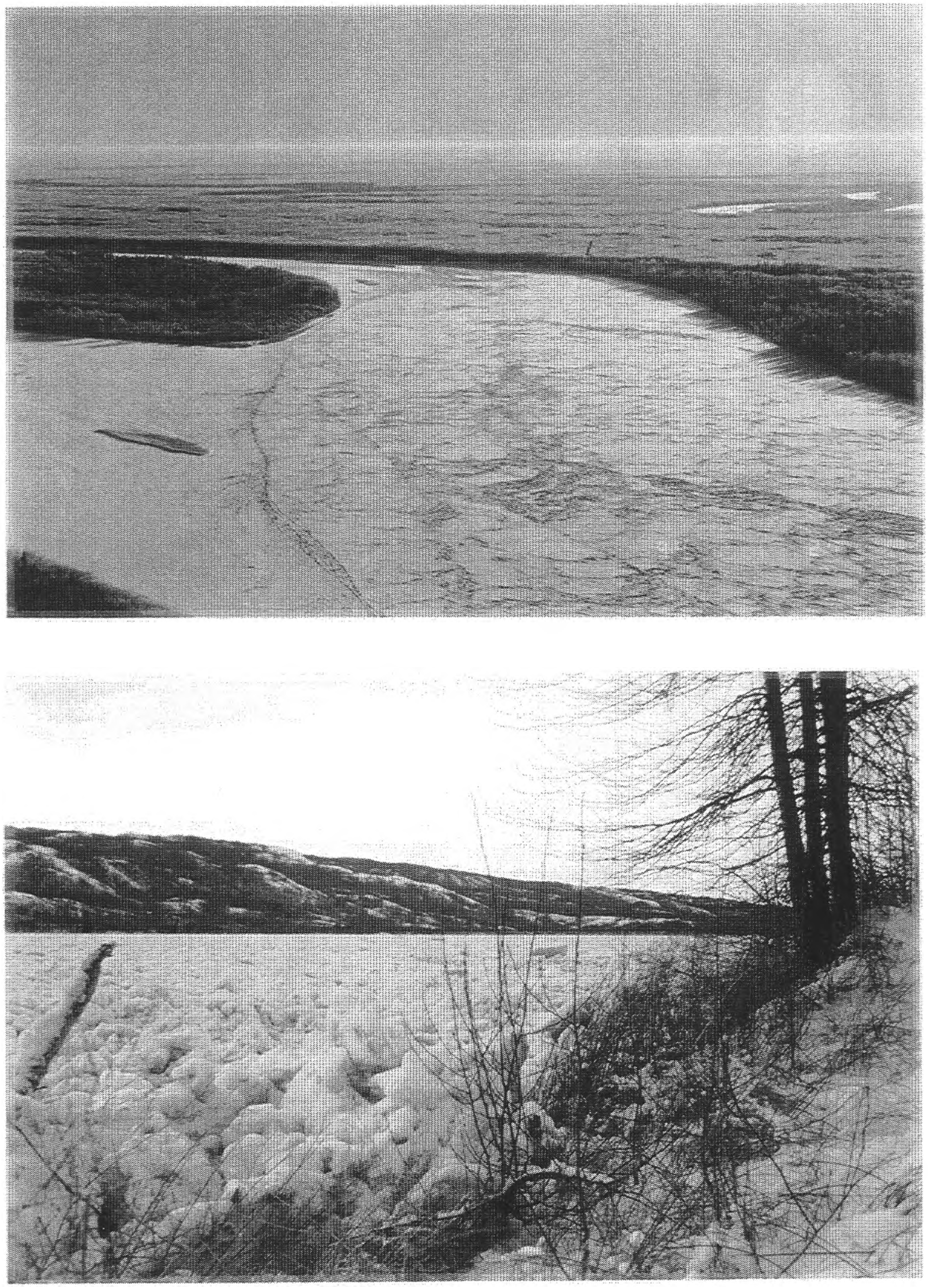


Figure 14 Examples of typical juxtaposed and consolidated ice covers.

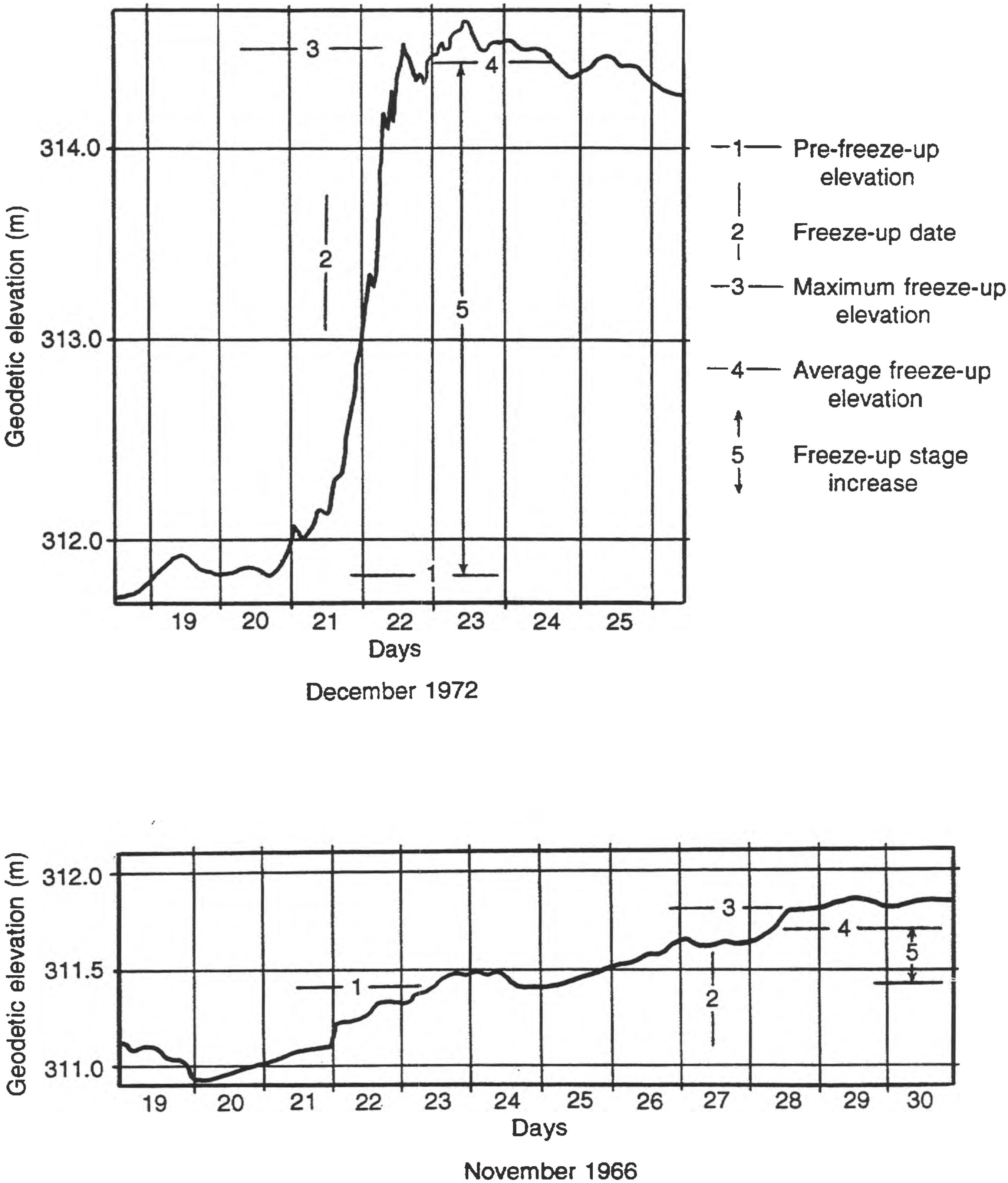


Figure 15 Typical gauge heights at freeze-up

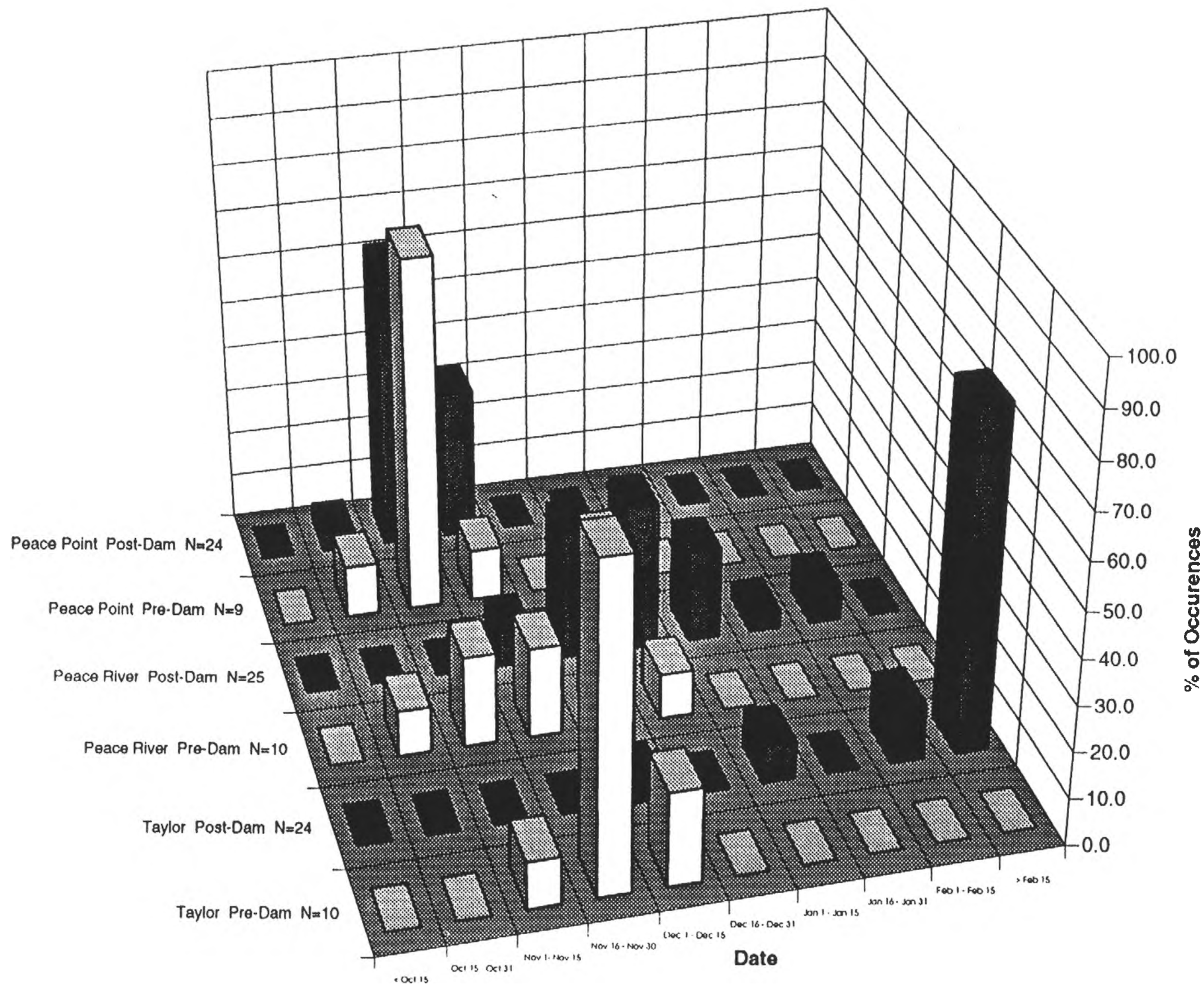


Figure 16 Comparison of pre- and post-regulation freeze-up dates

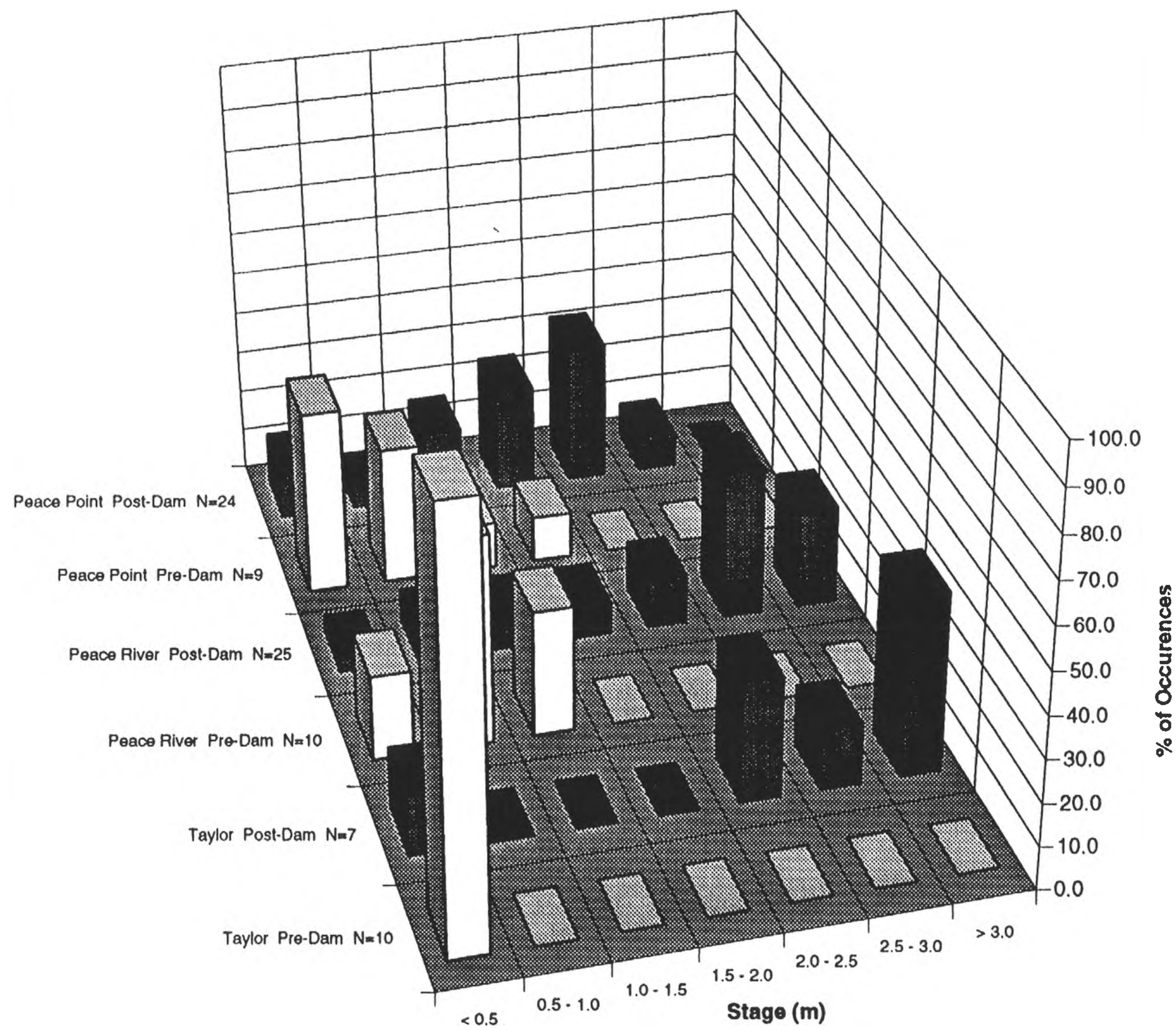


Figure 17 Comparison of pre- and post-regulation freeze-up stage increases

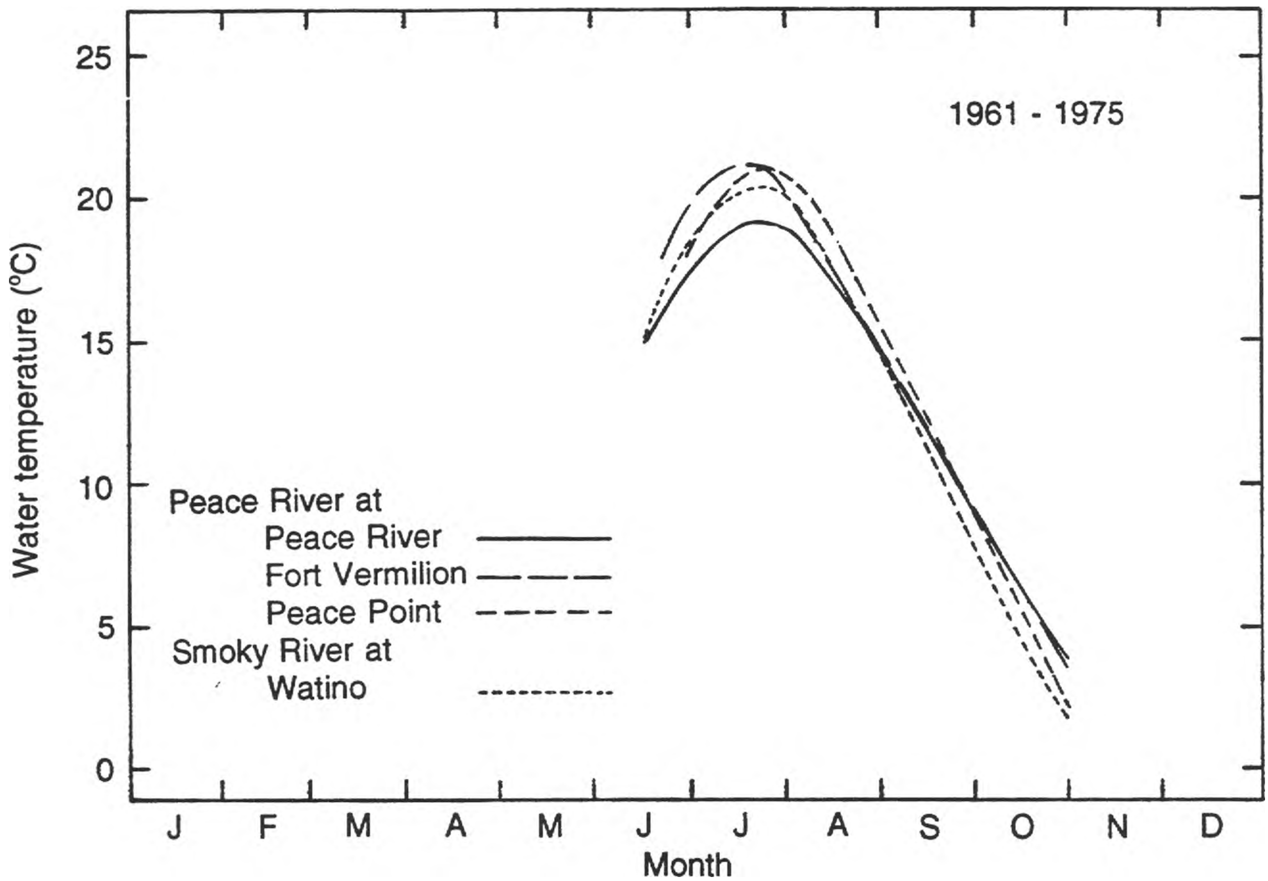


Figure 18 Measured water temperatures on the Peace and Smoky Rivers

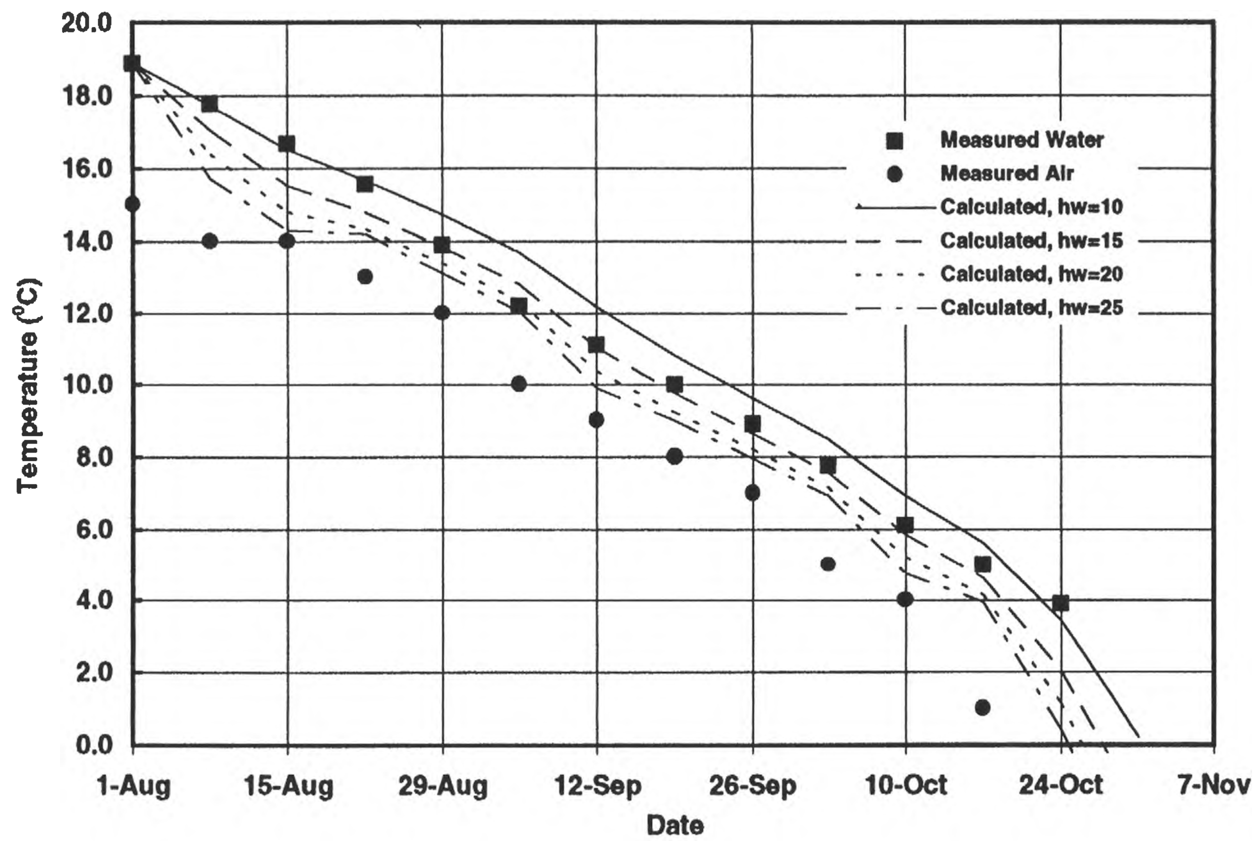


Figure 19 Comparison of measured and calculated water temperatures at Peace River prior to regulation

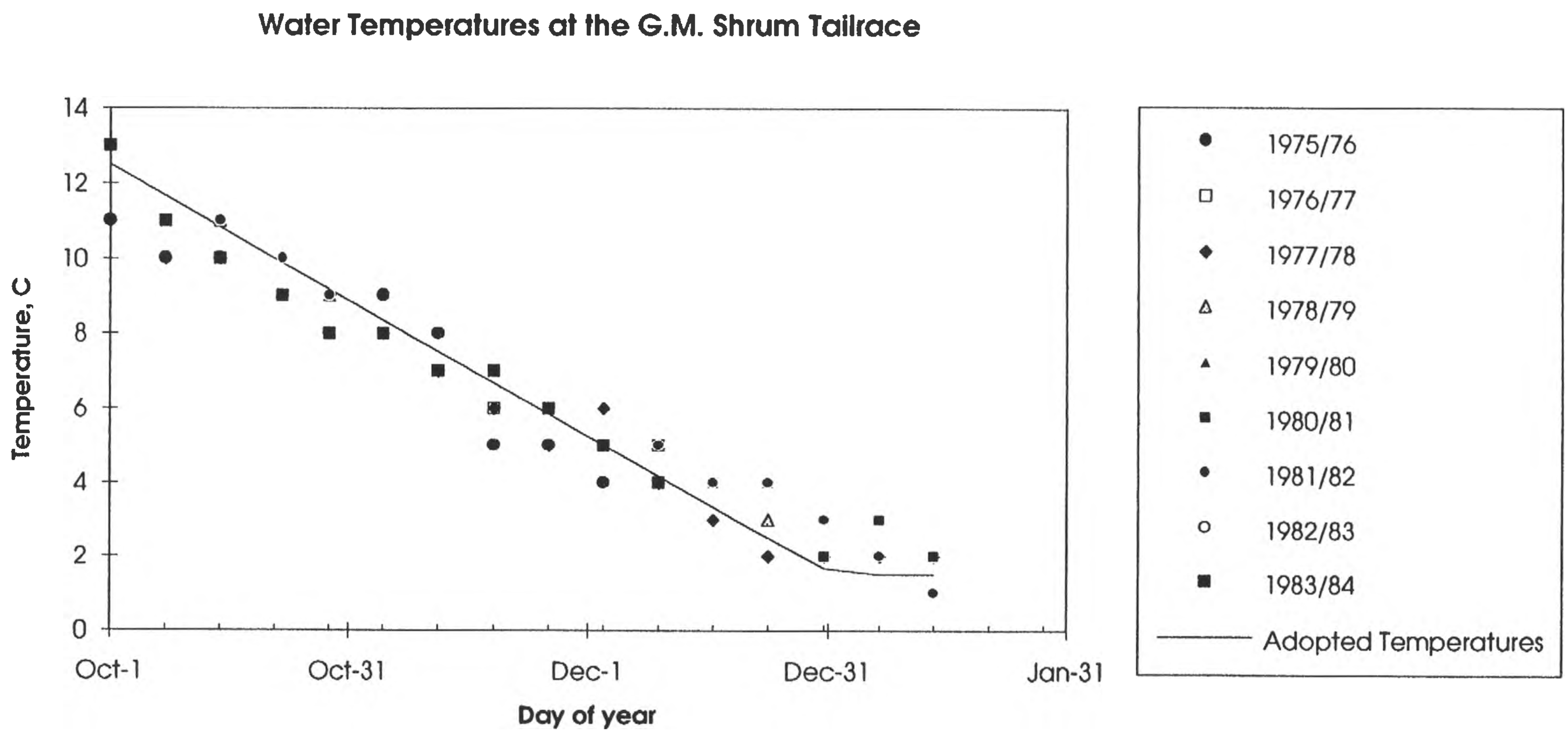
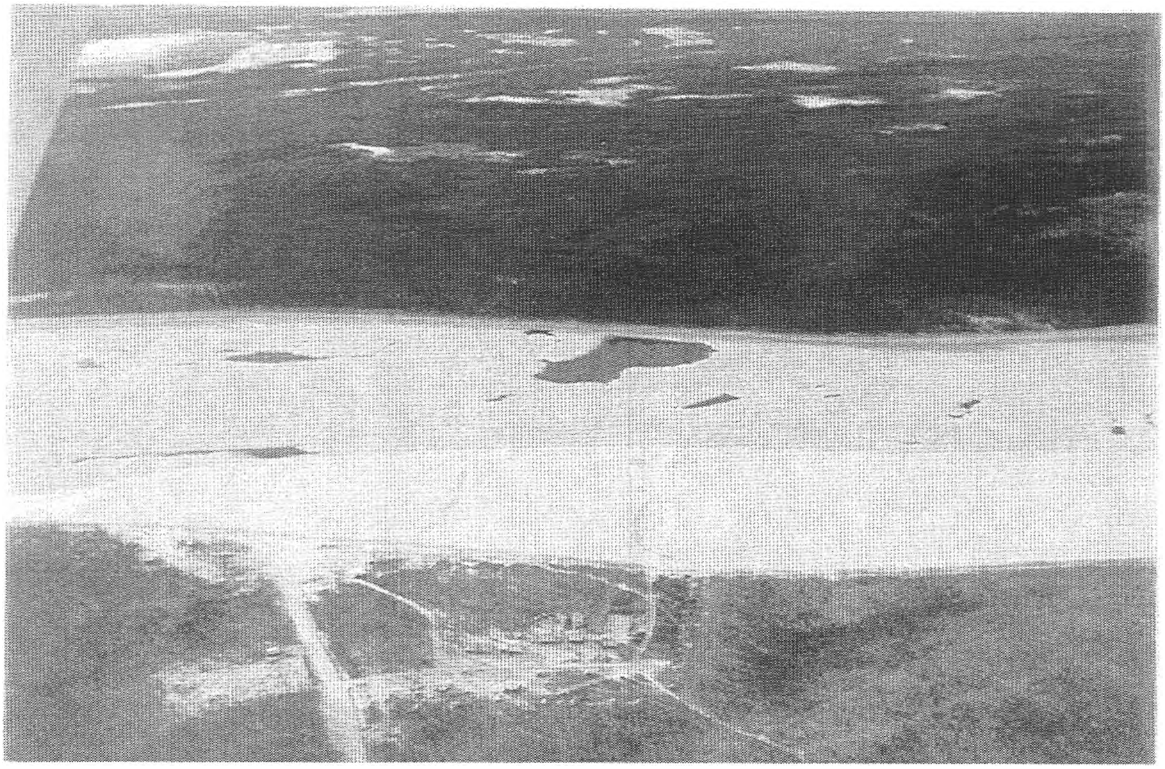


Figure 20 Variability in water temperatures downstream of the Bennett Dam during the freeze-up period



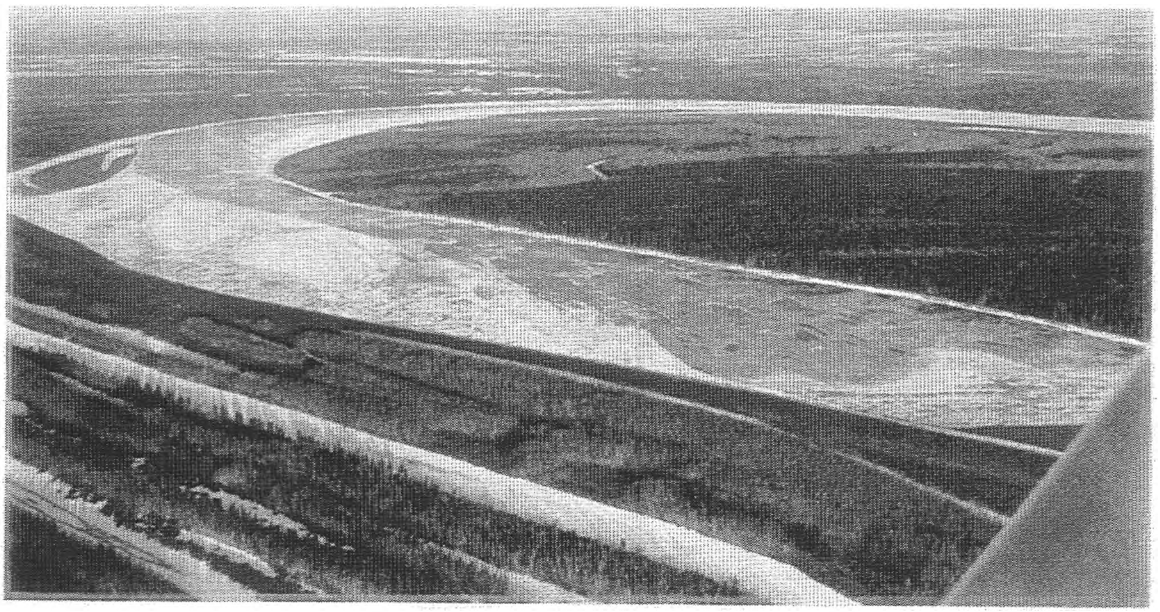


Figure 21 Typical ice characteristics after freeze-up downstream of Vermilion Chutes

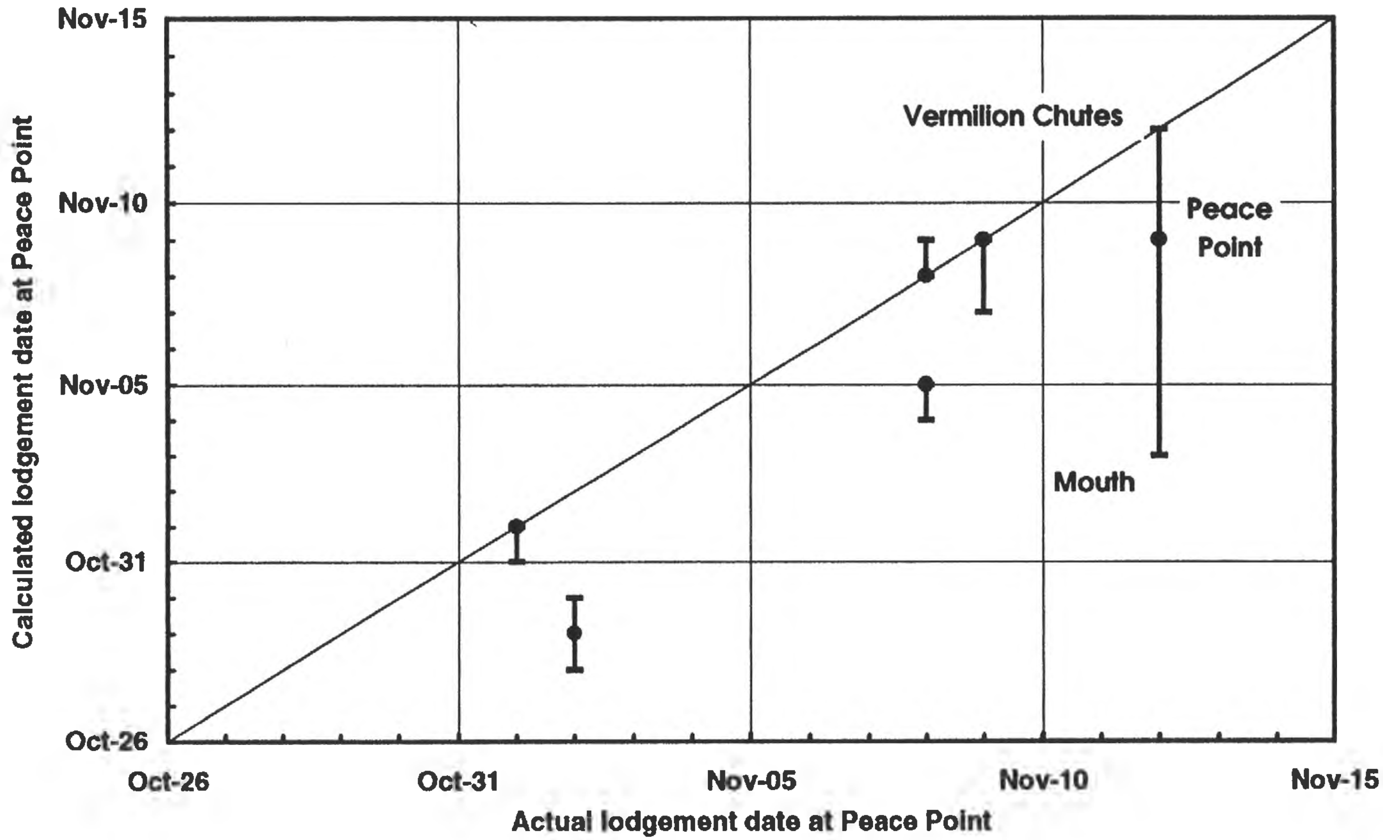


Figure 22 Comparison of the observed and simulated freeze-up dates at Peace Point after regulation

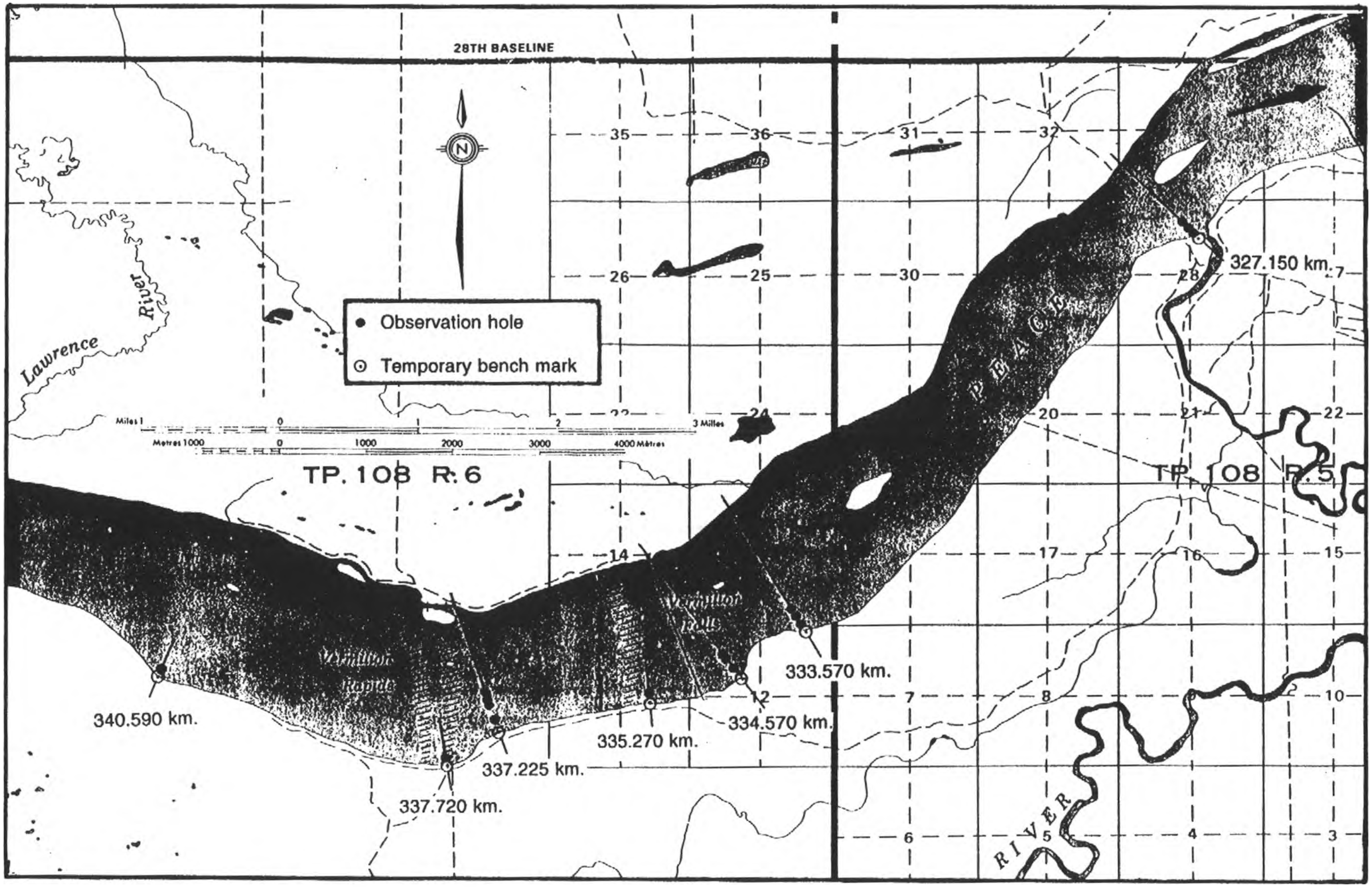


Figure 23 Vermilion Chutes study area

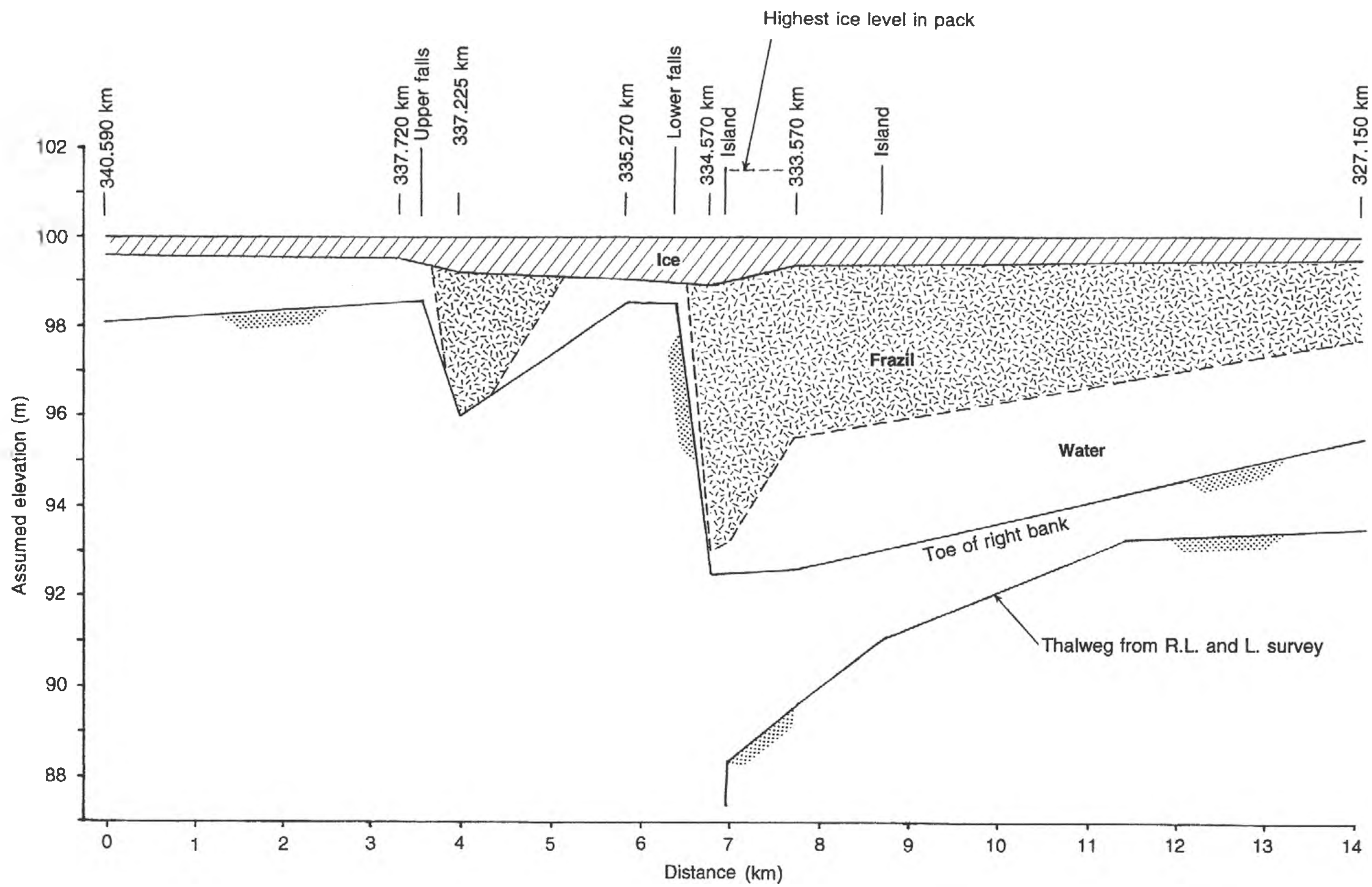


Figure 24 Slope profile through the Vermilion Chutes

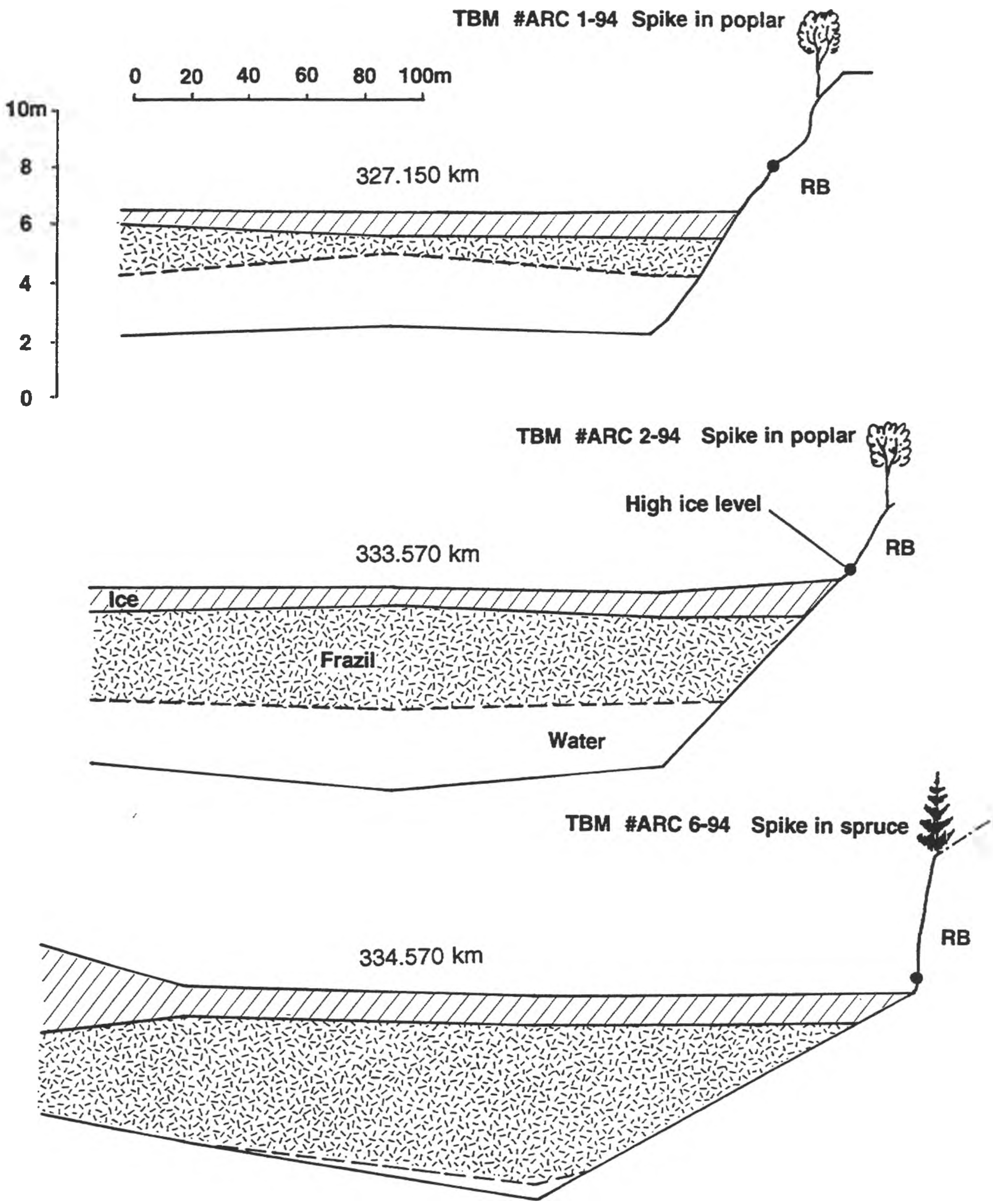


Figure 25 Ice thickness surveys in the vicinity of the Vermilion Chutes

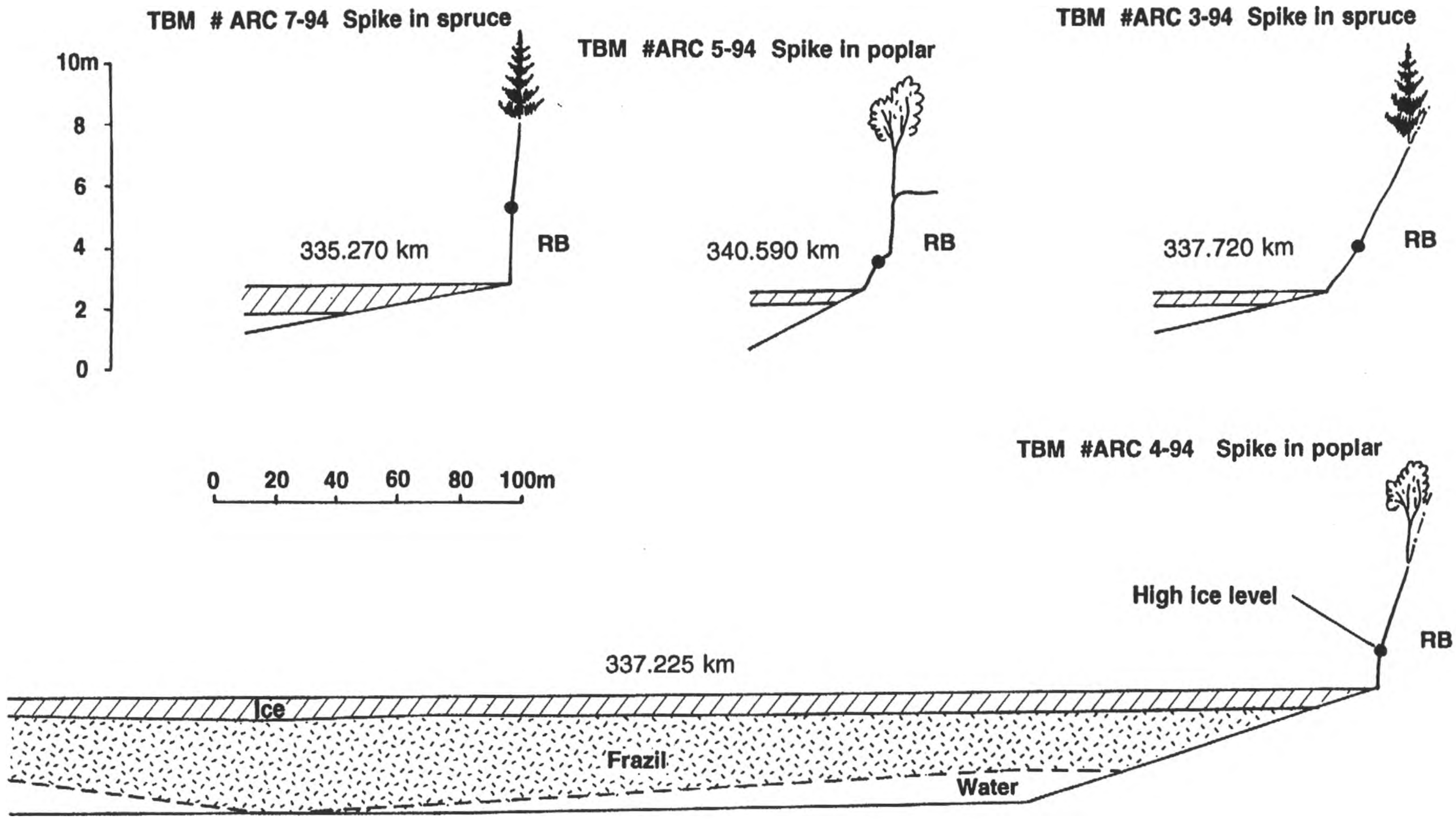


Figure 25 cont. Ice thickness surveys in the vicinity of Vermilion Chutes

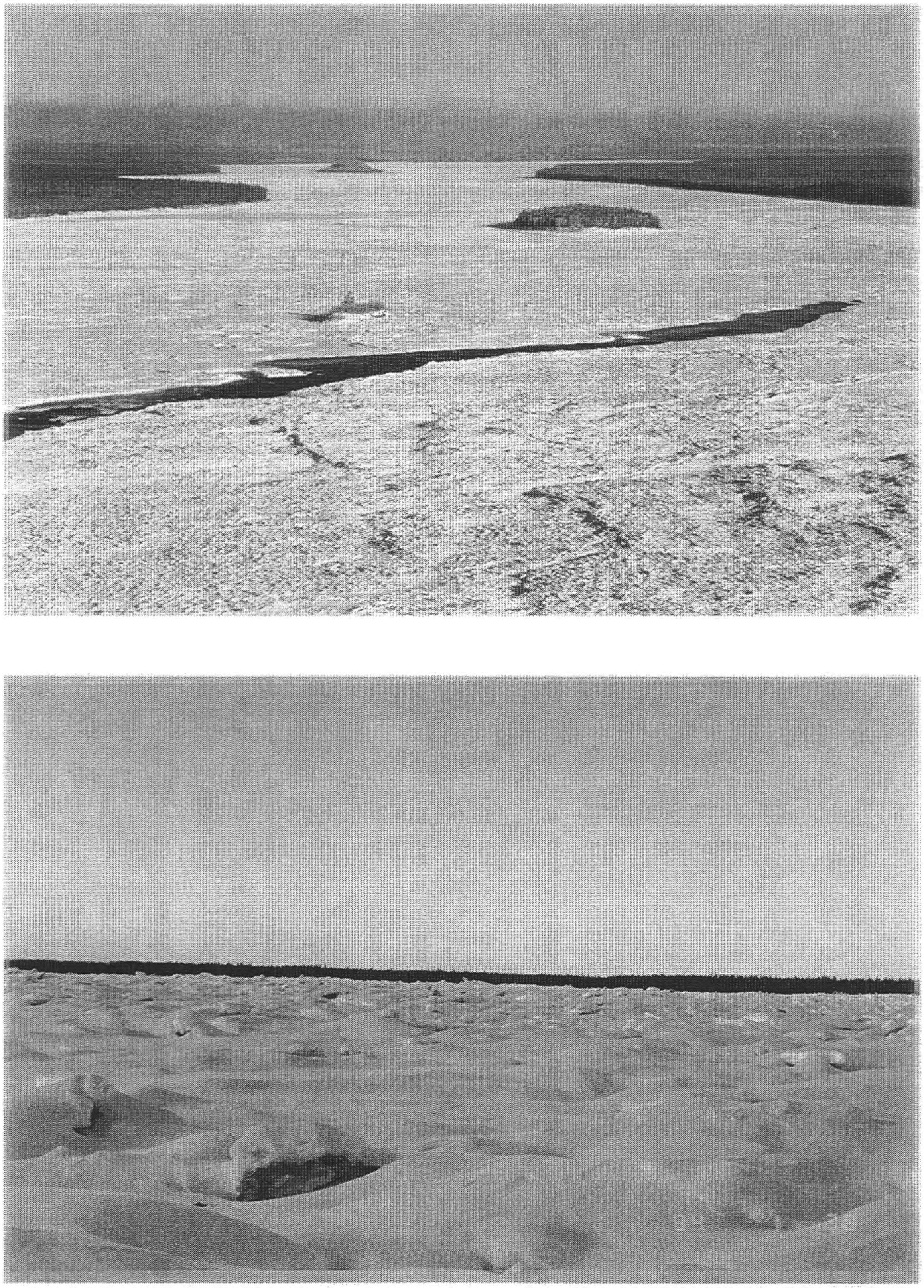


Figure 26 Characteristics of the ice cover in the vicinity of the Vermilion Chutes, 1993

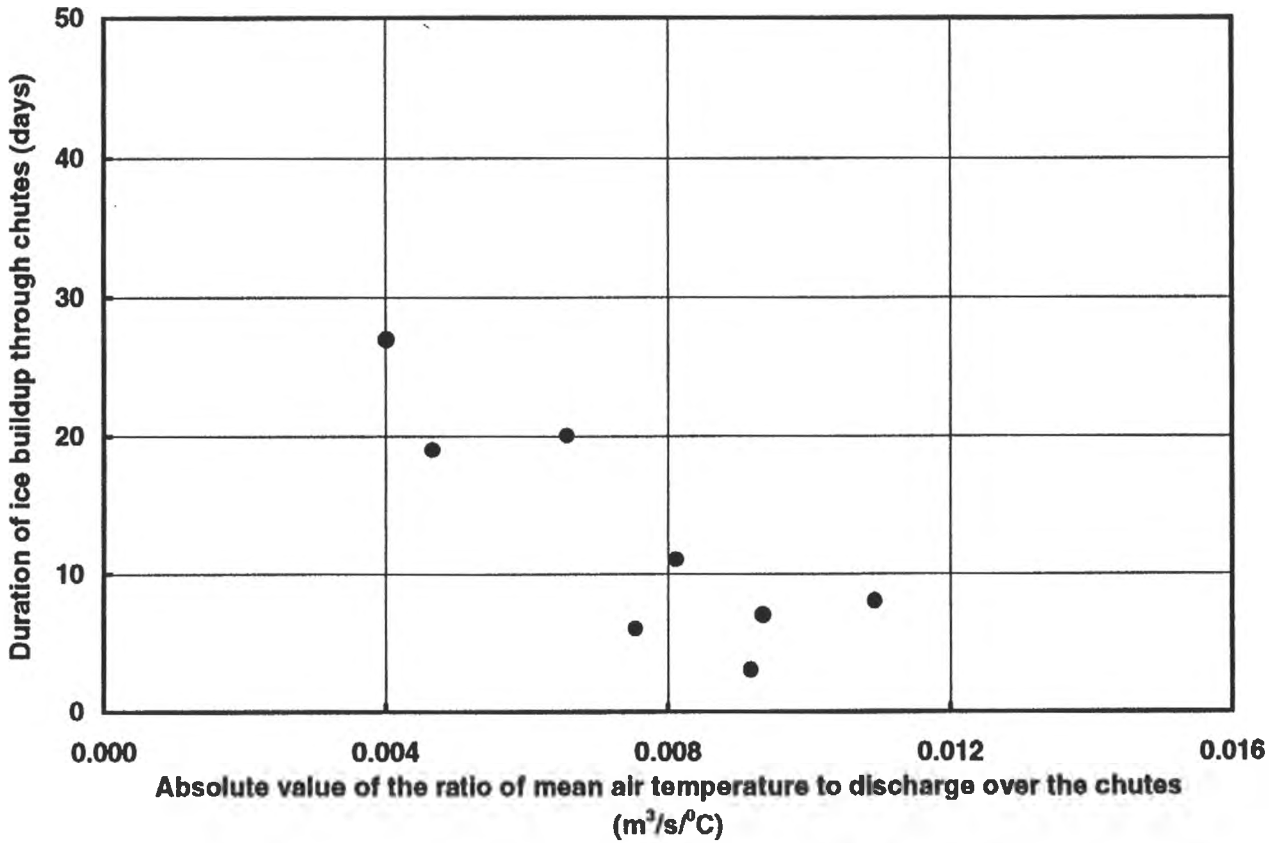


Figure 27 Criteria for the ice cover to stage over the Vermilion Chutes

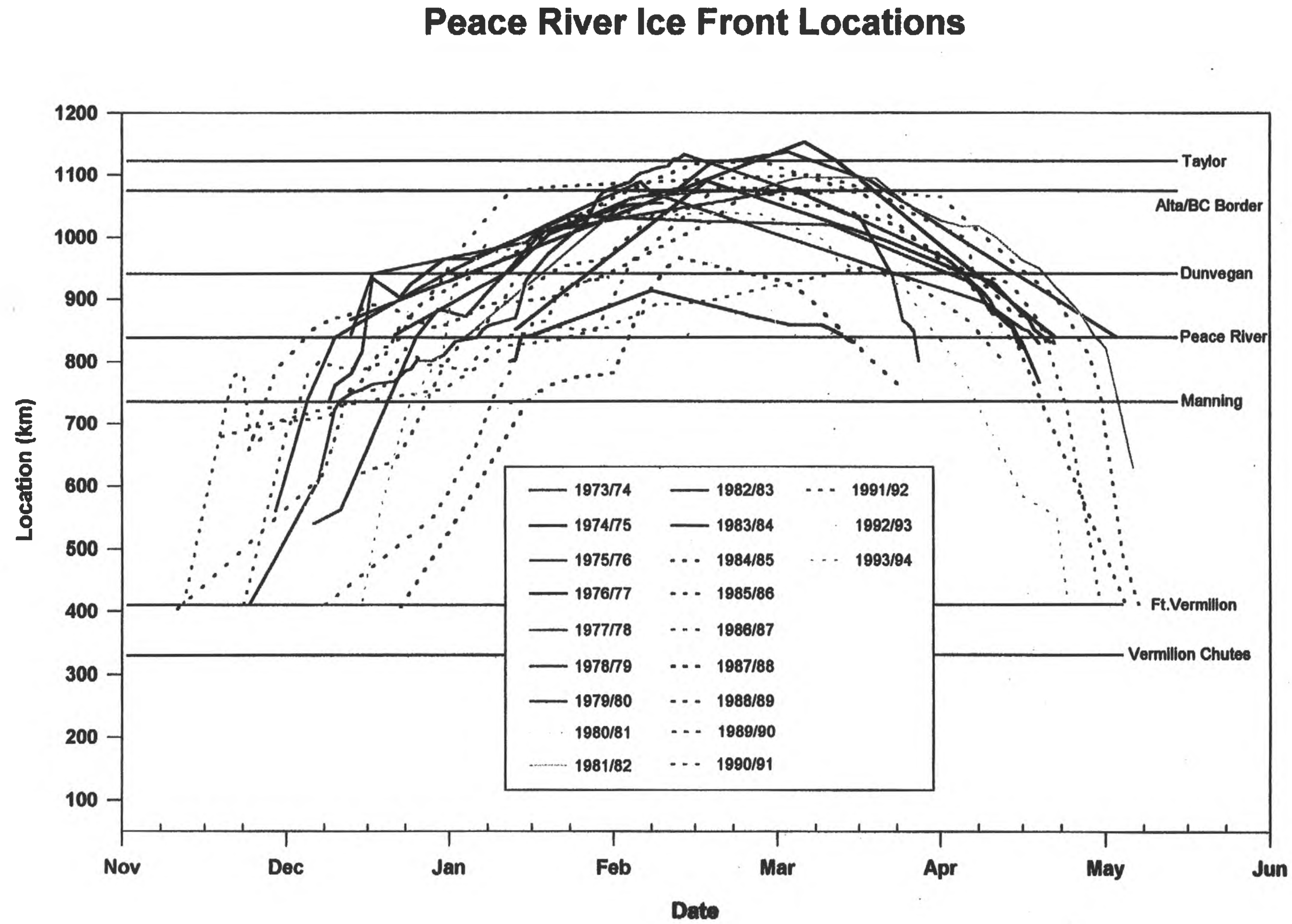


Figure 28 Ice cover progression rates after regulation, Fort Vermilion to Taylor

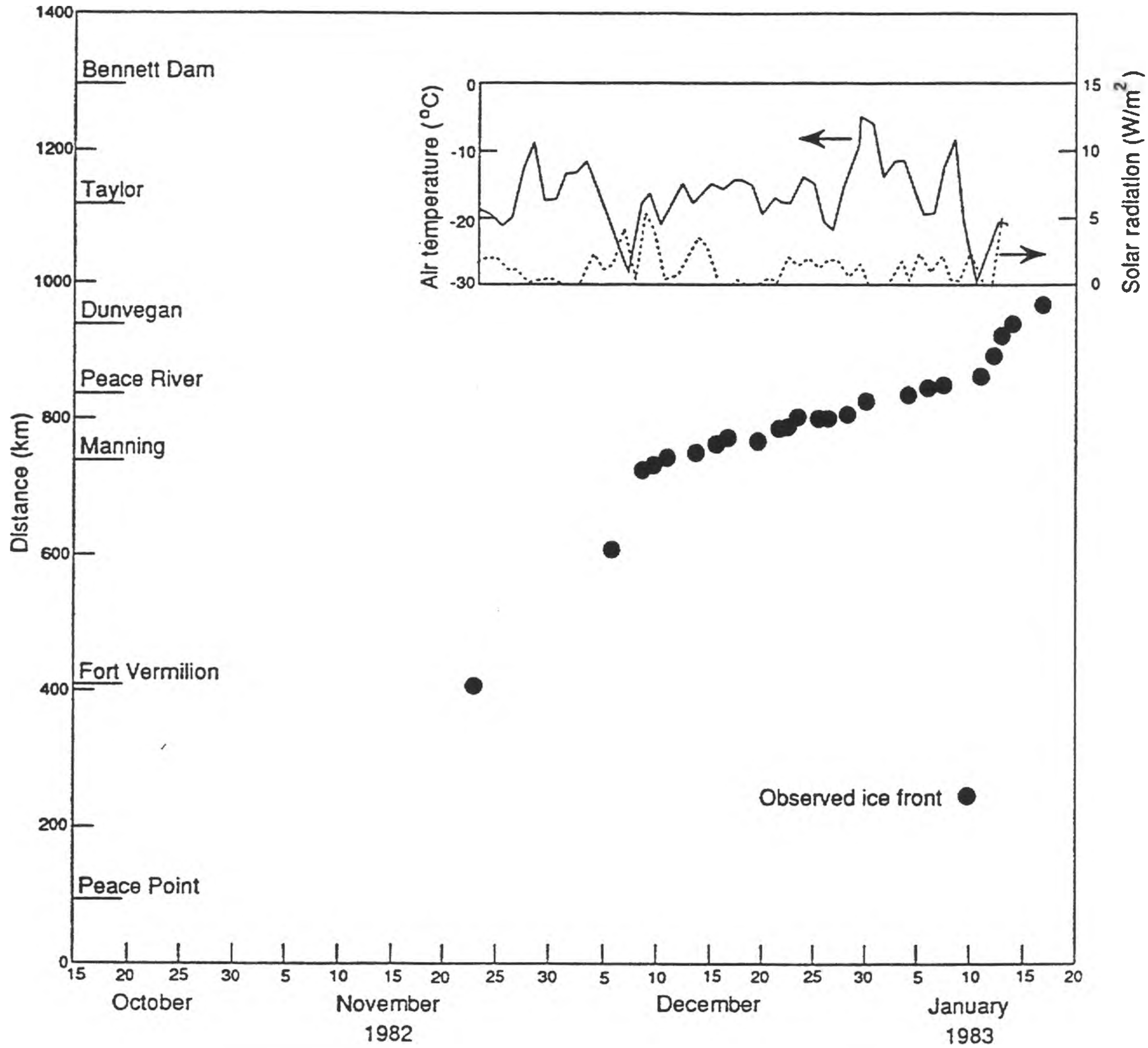


Figure 29 Ice cover progression rates during the 1982/83 freeze-up

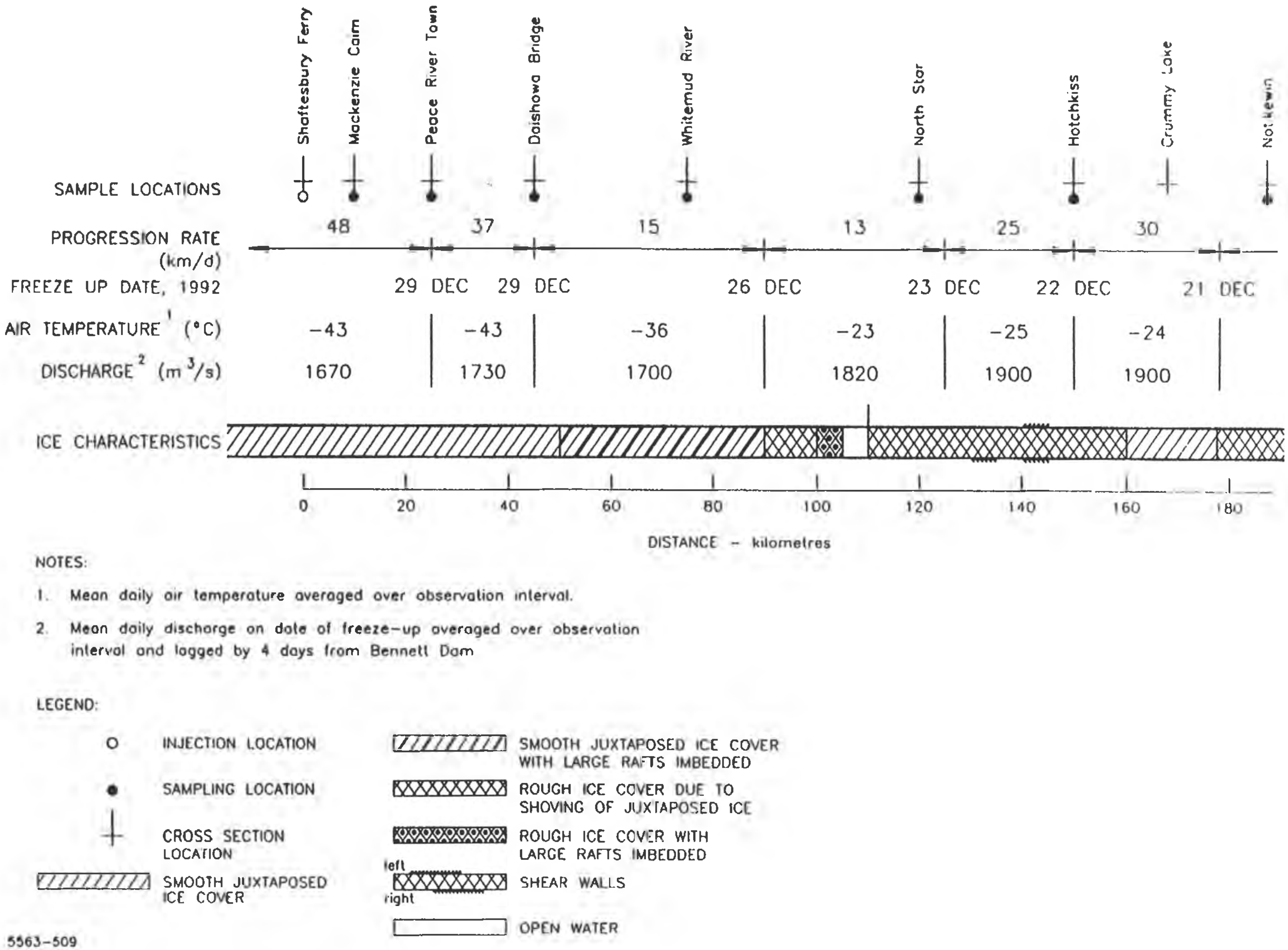


Figure 30 Ice cover formation characteristics, Notikewin River to Peace River, 1992/93

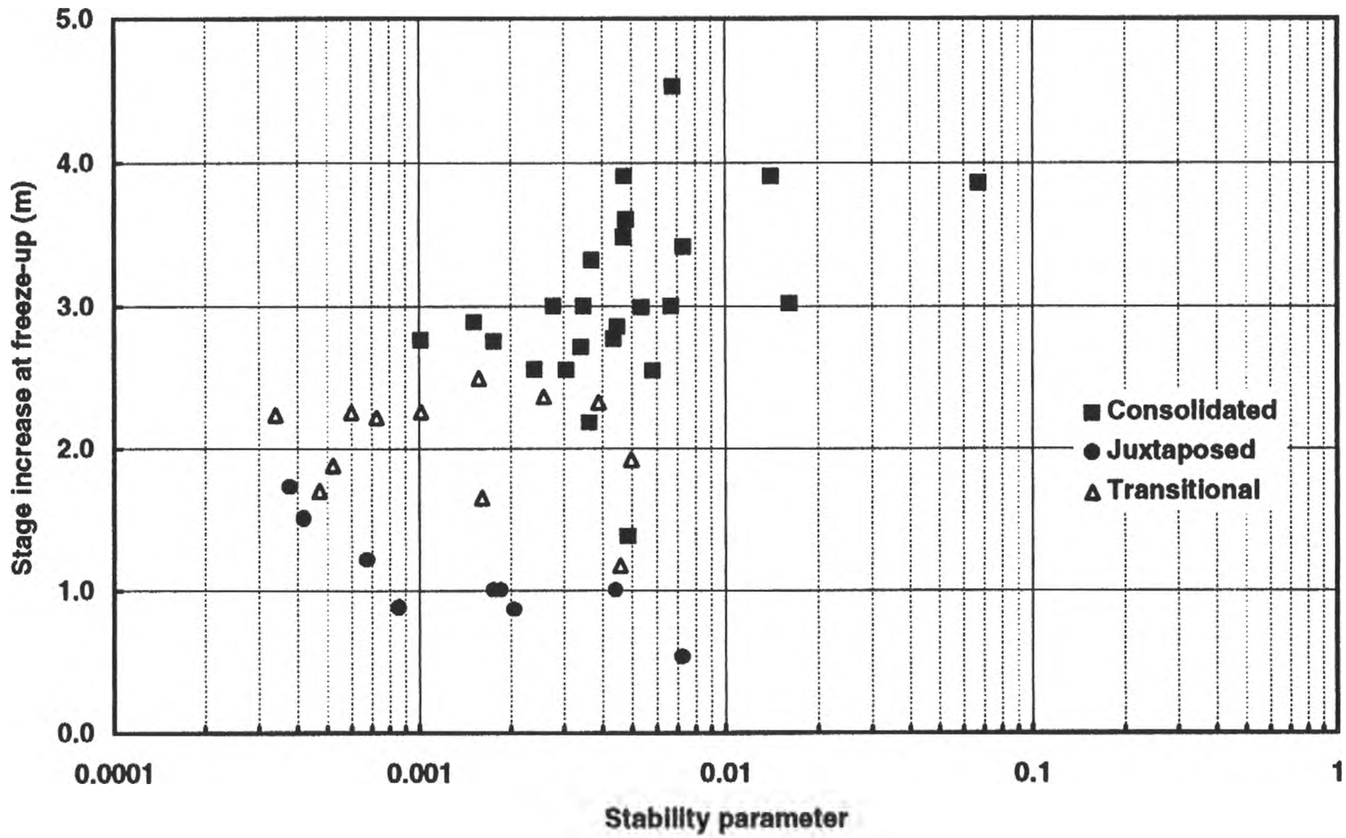


Figure 31 Stability criterion for juxtaposed ice covers under regulated conditions

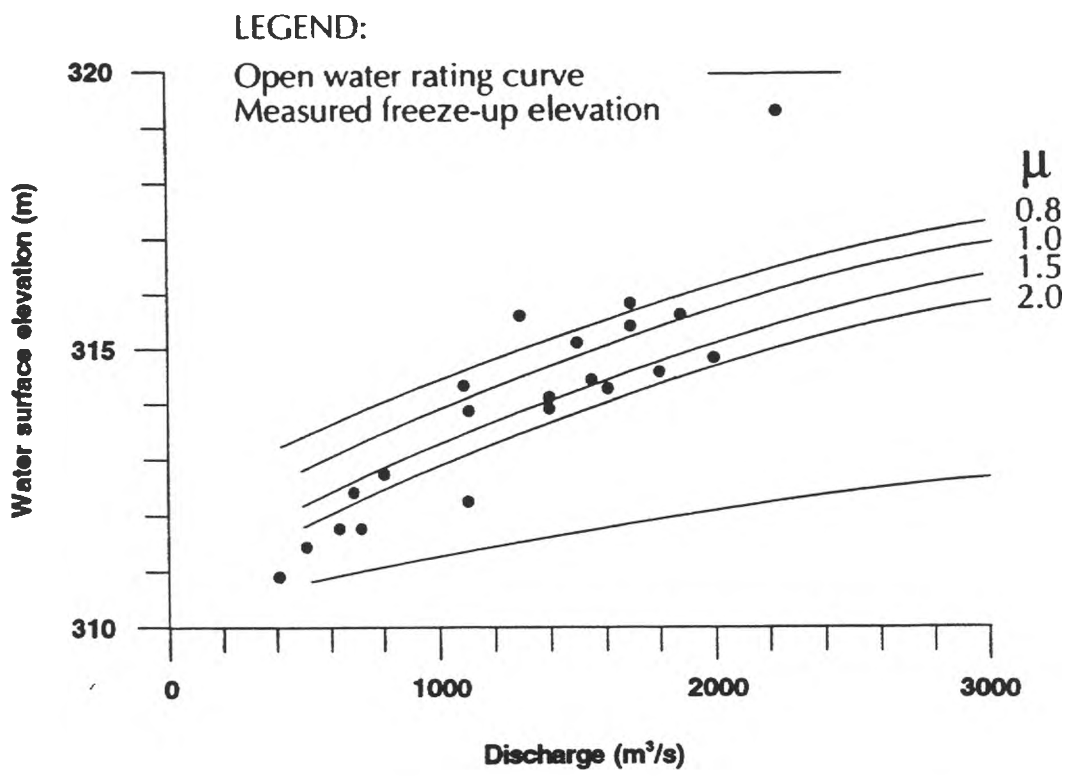


Figure 32 Calculated dimensionless internal friction coefficients for freeze-up ice covers at Peace River town

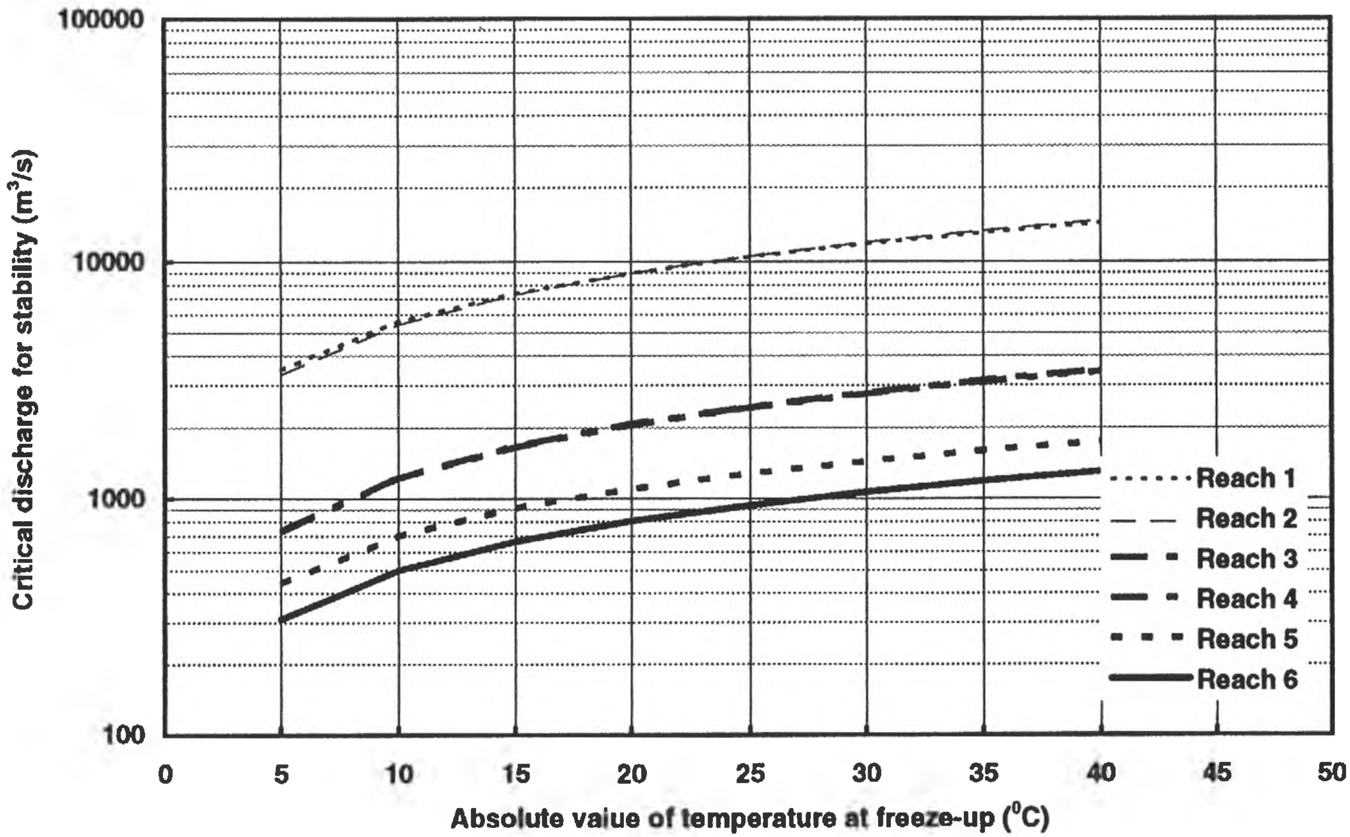


Figure 33 Ice cover stability as related to air temperature and discharge

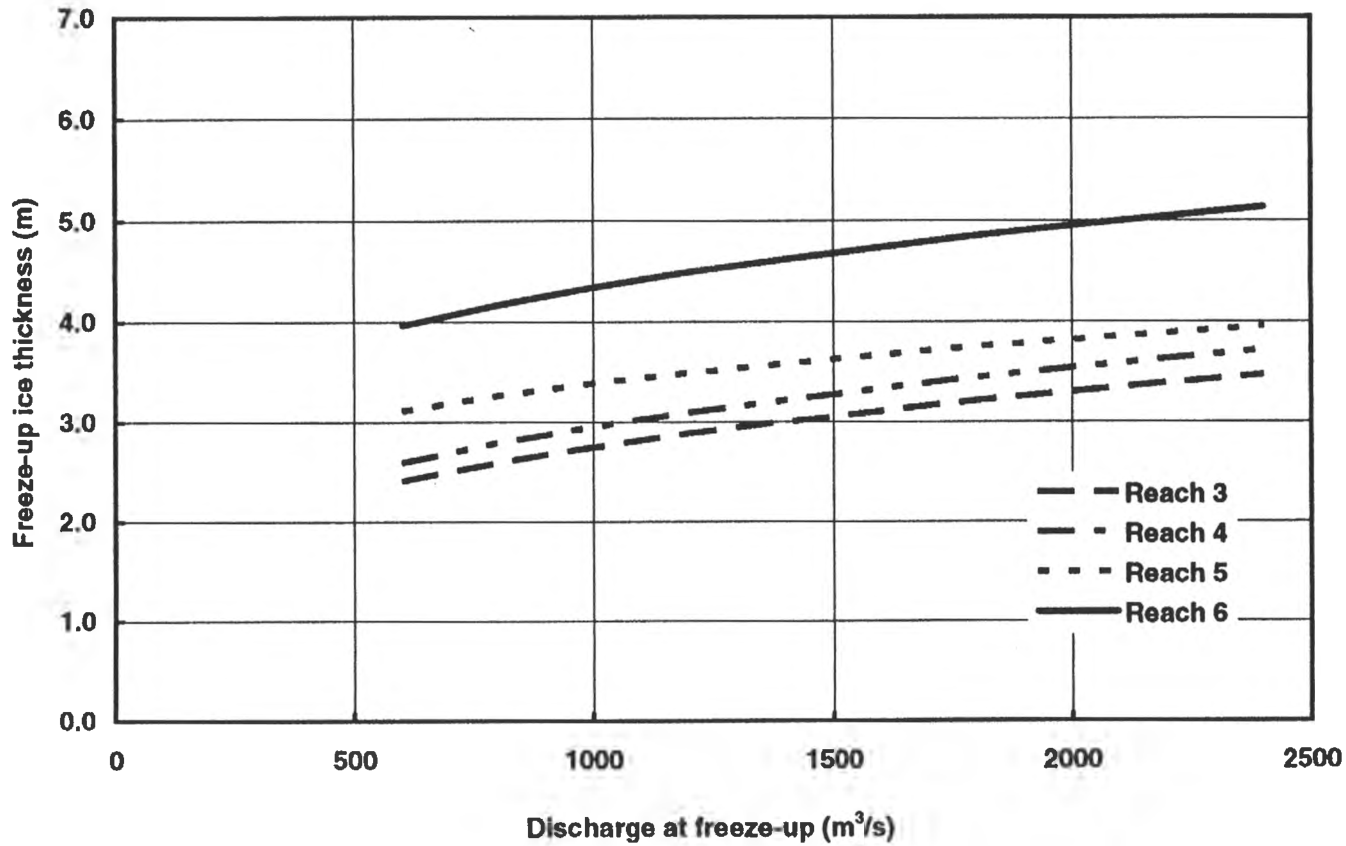


Figure 34 Ice cover thickness in a consolidated ice cover

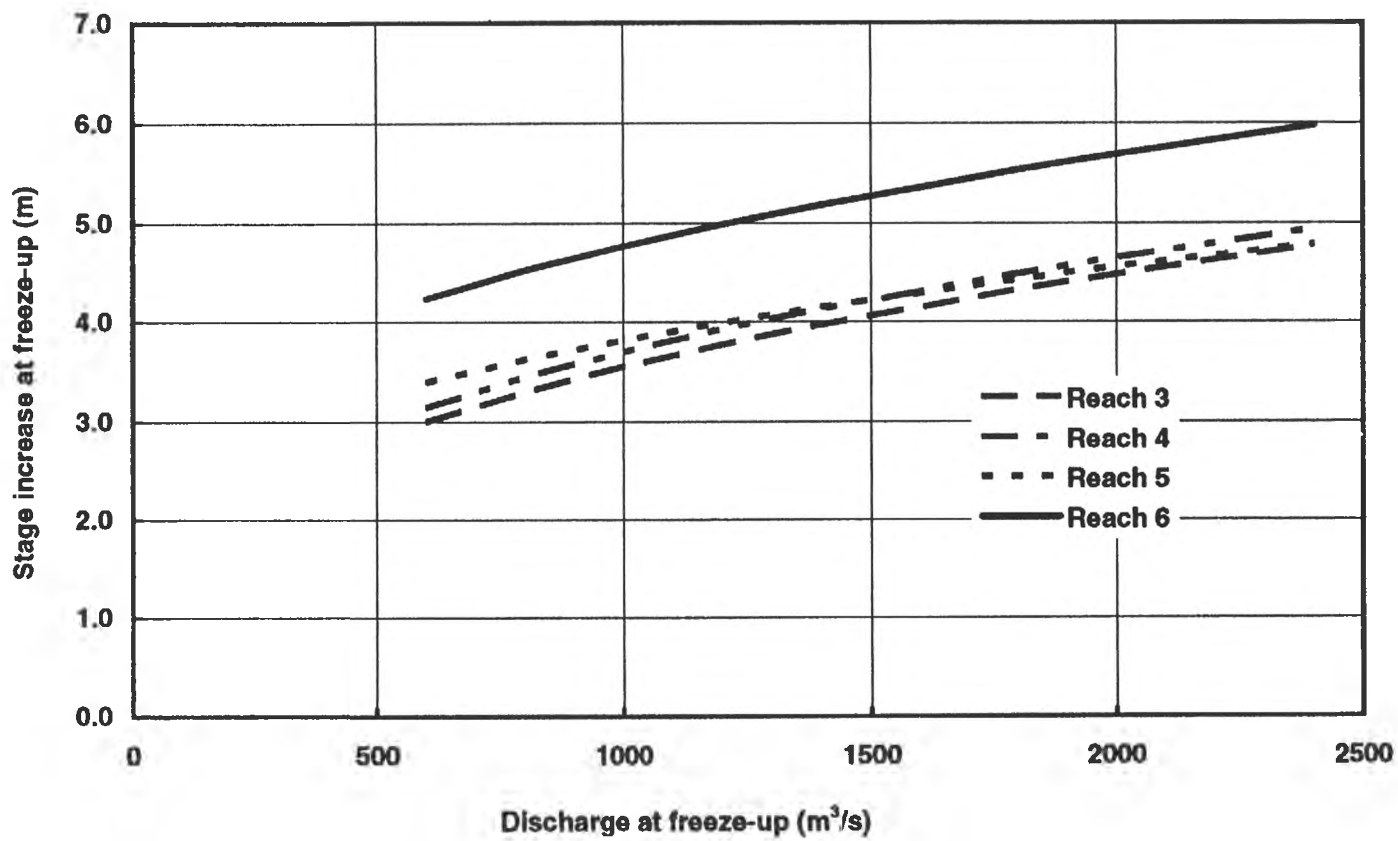


Figure 35 Freeze-up stage increase in a consolidated ice cover

APPENDIX A

TERMS OF REFERENCE

NORTHERN RIVER BASINS STUDY

SCHEDULE A - TERMS OF REFERENCE

Project 1423-C1: Freeze-up Characteristics on the Peace River, Taylor to the Slave River

I. Objective

The long term goals of this work are to quantify the importance of the freeze-up process on the hydrologic and biologic regimes of the Peace River. The work will identify the relevant processes which produce an ice cover on the river, compar the pre and post-dam freeze-up regime on a reach by reach basis, identify the impact of regulation on the characteristics and timing of freeze-up over the reach of interest, and determine the boundary and initial conditions for the subsequent (or concurrent) analysis of the breakup process. This project will combine the various ice observations, hydrometric records, and hydraulic characteristics to identify nature of the ultimate hydrologic and biologic regimes of the Peace River and the Peace/Athabasca delta.

II. Requirements

1. The study area will be described with respect to its location, physiography and political boundaries.

2. The pre and post-dam flow will be identified for each freeze-up period for each year. Dr. Faye Hicks of the University of Alberta and Mr. John Taggart of Alberta Environmental Protection, will be consulted to determine the pre and post-dam flows and the naturalized flows in the post dam period.

3. The hydraulic characteristics of the relevant identifiable reaches will be summarized and their sensitivity to changes I the freeze-up discharge regime will be determined.

4. The freeze-up characteristics (date, thickness, ice cover type) will be summarized for each of the hydrometric gauges and compared for the pre and post-dam periods.

5. Dominant freeze-up modes (as a function of temperature and discharge) will be identified for each of the reaches. The effects of regulation on the ultimate character of the dominant ice cover will then be assessed.

6. Attempts will be made to extend the freeze-up characteristics to issues related to ground water levels and habitat characteristics adjacent to the river.

7. A report will be produced to document the above work.

III. Project Organization

this project will be managed by Mr. Dave Andres of Alberta Research Council. Scientific collaborators include Mr. Gary Van Der Vinne, Mr. J. Thompson, and Mr. P. Mostert, all of the Alberta Research Council.

IV. Reporting Requirements

As per NRBS