### Table 9.2 - (CONTINUED)

<table>
<thead>
<tr>
<th>DESCRIPTION OF FIELD DATA</th>
<th>ICE PROCESS</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(A)</td>
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<td></td>
<td>Acoustic sounding</td>
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<td>9.0 OBSERVATIONS OF FORMATION, PROGRESSION AND REGRESSION OF ICE COVERS</td>
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<td>Ice cover occurring studies</td>
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<td>Satellite photos</td>
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<td>Ground observations</td>
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<td>10.0 COVER TYPES, STRENGTHS AND TOUGHNESS</td>
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<td>Snow cover</td>
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<tr>
<td>Slush ice/ice slush</td>
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<tr>
<td>Ice cover, fixed</td>
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<tr>
<td>Ice cover, drifting</td>
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<tr>
<td>Ice cover, breaking</td>
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<tr>
<td>Ice cover, covering</td>
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<td>11.0 EVOLUTION OF FREEZE-UP ICE JAMS</td>
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<tr>
<td>Initial stage</td>
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</tr>
<tr>
<td>Collapse and relocation</td>
<td>1</td>
</tr>
<tr>
<td>Role of sliding and mechanical erosion</td>
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</tr>
<tr>
<td>Ice jet failure</td>
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</tr>
<tr>
<td>12.0 EVOLUTION OF FREEZE-UP ICE JAMS</td>
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</tr>
<tr>
<td>Appearance of ice cover</td>
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<tr>
<td>Observed ice cover</td>
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<tr>
<td>Formation of the breach</td>
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<tr>
<td>Initial stage of ice jam break-up</td>
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<tr>
<td>13.0 EVOLUTION OF SPRING BREAKUP ICE JAMS</td>
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<tr>
<td>Breaching</td>
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<tr>
<td>Collapse and relocation</td>
<td>1</td>
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<tr>
<td>Role of sliding and mechanical erosion</td>
<td>1</td>
</tr>
<tr>
<td>Ice jet failure</td>
<td>1</td>
</tr>
</tbody>
</table>

**NOTATION:**
- 1: Very Important
- 2: Moderately important (could be very important on certain rivers)

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**Chapter 9**

**A CASE STUDY: FREEZE-UP AND JAMMING ON THE PEACE RIVER AT PEACE RIVER**

David D. Andres

9.1. INTRODUCTION

The Town of Peace River is located on the banks of the Peace River below the confluence of the Smoky River, about 400 km downstream of the W.A.C. Bennett Dam (Fig. 9.1). Since the regulation of the Peace River by the dam in 1972, the winter flows at Peace River town have been increased to approximately four times the previous natural flows. This has altered considerably the ice conditions in the river. Most notably, the freeze-up date at the town has been delayed from early November, under the natural flow regime, to December or early January and freeze-up levels have increased substantially. This has resulted in near flood conditions on a number of occasions and heightened the public awareness of the impact of hydropower operations on river ice conditions.

Considerable freeze-up information has been gathered on the Peace River at the town (Andres, 1978; Acres Consulting Services Ltd., 1980, 1984; Carson and Lavender, 1980; Fonstad, 1982, 1986; Fonstad and Garner, 1984, 1986; Neill and Andres, 1984; and Biberhofer, 1984). The data includes annual observations of ice processes, Water Survey of Canada (WSC) water level records and interpretations thereof, and detailed measurements of the ice cover characteristics for particularly severe freeze-up conditions in 1981/82.

The purpose of this chapter is to provide a case study which describes and quantifies freeze-up processes and ice jams on a large river affected by hydropower regulation. A well documented event during the 1981/82 freeze-up is analyzed to define the hydraulic roughness typical of the ice cover at freeze-up. With this result, it is then possible to generalize and further elucidate the more limited Water Survey of Canada (WSC) data which have
been gathered over a period of about 30 years. This will illustrate the application of some of the available methods to analyze ice jam situations of this type and ultimately to establish design criteria for their management.

9.2. BACKGROUND

The Peace River is situated in northern British Columbia and Alberta. It 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.

The climate of the Peace River region is essentially the humid micro thermal, sub-arctic or cold “snowy forest” type. The exception is the area between Taylor and Peace River, which is classified as humid continental, with cool summers and no dry season (Longley, 1972). The November, December, January normal mean monthly temperature at Peace River is -8.1, -15.3, and -10.4°C, respectively. The precipitation during those months is 20.0, 21.6, and 22.1 mm, respectively.

The Peace River has been gauged at Peace River since 1915 with a gap from 1932 to 1957. The annual mean flow is approximately 1800 m³/s. Winter flows under natural conditions were mostly in the range of 200 to 500 m³/s, but under regulated conditions since 1972 these have been increased to between 1000 to 2000 m³/s (Fig. 9.2).

![Monthly flows on the Peace River downstream of Peace River](image)

The Peace River, within the study area near Peace River town, flows in a 200 m deep stream-cut post glacial valley with a top width of about 5 km. The bottom width of the valley is about 1300 m and exhibits narrow fragmentary terraces. The channel is sinuous, with occasional islands and mid channel bars, and is partly entrenched and confined. The channel bed consists of about 5 m of gravel over soft shale. The channel banks are composed of moderately erodible shales, overlain by gravel and silt. At high river stages near bankfull, the channel width averages about 550 m and the mean depth is about 6 m. The water surface slope is 0.29 m/km. Figure 9.2 illustrates the channel pattern and its relationship to the valley.

Figure 9.4 shows the open-water rating curve for the range of discharge anticipated during freeze-up. Also included in the figure is the first winter measurement (made some 10 to 30 days after freeze-up) following the formation of a stable ice cover. The additional backwater due to the thickness
A Case Study: Freeze-up and Jamming on the Peace River at Peace River

Fig. 9.4. Open water and early winter rating curves on the Peace River at Peace River.

of the ice and its additional roughness is evident. This backwater increases with increasing discharge indicating at least a thicker, and perhaps rougher, ice cover at the higher freeze-up discharges following regulation.

Figure 9.5 illustrates the reach-average hydraulic geometry of the Peace River within the study area (Fig. 9.6). Both the flow area and the mean depth vary linearly with stage as represented by the gauge elevation. This is due to

Fig. 9.5. Reach-averaged hydraulic characteristics of the Peace River at Peace River.
A Case Study: Freeze-up and Jamming on the Peace River at Peace River

the more or less constant top width above elevation 312 m. The reach-
averaged open-water Manning roughness coefficient of the bed varies between
0.045 at a mean stage of 2 m (Q = 600 m³/s) to about 0.025 for a mean stage
greater than 4 m (3600 m³/s). At typical freeze-up discharges of 1500 m³/s the
open-water bed coefficient is less than 0.030.

A geomorphic anomaly exists in the river channel between the mouth of the
Heart River and the head of Belly Island. The channel here is about twice as
deep as that upstream with a similarly larger flow area at any given discharge.
The WSC gauge is situated at the downstream end of this anomaly and most of
the open water and ice covered discharge measurements and ice thickness
measurements have been made in its deepest part. This has biased somewhat the
"at a section" hydraulic characteristics gathered for the river at this gauge
and they are not entirely representative of the reach of interest.

9.3. FREEZE-UP PROCESSES

The temperature of the water released from Williston Lake is in the order of
1°C to 4°C during freeze-up. The released water cools as it moves downstream,
frazil is generated, floats to the top, and the surface concentration of the ice
approaches unity over a relatively short reach. This limits the production of
frazil to typically less than half of what one might expect for completely open
water conditions. Mathematical modelling (Andres and Spitzer, 1989)
suggests a porosity of 0.7 for the frazil attached to the floes and a slush
thickness of 0.5 m once 100% coverage is achieved. Border ice growth is
limited to back channels and low-velocity, shallow areas, thus producing a
relatively uniform width of open water. However, no observations have been
made to quantify this process.

Downstream of Peace River, initial cover formation seems to occur by
bridging near Fort Vermilion. The cover progresses upstream by a
combination of juxtaposition first (surface jam), followed by shoving. Initially,
the ice cover progresses by the juxtaposition of individual floes or large rafts
which are too large to be submerged at the head of the cover. This produces a
thin cover which gains strength from downward freezing. Shoving will occur
if the rate of incoming ice is sufficient to lengthen the cover rapidly enough so
as to generate sufficient downstream force (by the development of shear on the
underside of the cover) in order to overcome the resisting force that is being
developed by downward freezing in the thin accumulation.

Downstream of Peace River, where the river slope is in the order of 0.05
m/km, the cover may or may not shove, depending on a delicate balance
between the development of the driving force and the resisting force. At Peace
Chapter 9

River and upstream, the river slope is an order of magnitude greater, thus almost always ensuring snowing because the steeper slope causes much larger shear stresses on the ice cover and there seldom is sufficient time for significant resisting forces to develop.

In certain situations, such as when upstream ice production is limited by either warm temperatures or by low winds, juxtaposition is the primary mode of permanent cover formation. This situation may be enhanced by a gradient of decreasing air temperatures from west to east and south to north. This temperature gradient results in limited ice production in the ice generation zone and yet rapid downward freezing in the zone where the ice cover is forming. The low discharge limits the amount of ice being produced and also significantly reduces the rate at which force is applied on the juxtaposed cover.

This type of phenomenon results in a rather characteristic record of stage (Fig. 9.7) at the WSC gauge. Assuming a constant incoming discharge, the water level increases (1) as backwater is felt at the gauge from the formation of the ice cover downstream. A temporary, unstable, high freeze-up level (2) due to the presence of the juxtaposed limb is followed by another increase in stage (3) resulting from shoving. This stage increase is due to both the increasing ice thickness and the increasing discharge as flow is released from channel storage when the ice cover collapses. Depending on the extent of the shoving, open water may or may not develop at the gauge but the backwater is usually evident. Nevertheless, the final configuration is usually a stable ice cover (5) at a level somewhat less than the peak stage (4) because of the reduced discharge due to the passage of the flood wave associated with the shoving and because a part of the flow is being abstracted and placed into channel storage as the cover progresses upstream.

Very often the above process occurs very quickly or even two or three times over the freeze-up period. This can result in a very complicated gauge record that is difficult to interpret. The main difficulties are estimating the peak discharge associated with the final shoving event (4), the actual flow at the gauge during the stable cover period (5), and the thickness and roughness of the ice cover.

Given the similarities in each year’s freeze-up events, it can be reasonably assumed that the resulting ice cover probably has the same underside characteristics and hence the same hydraulic roughness each year, even though the thickness may be considerably different. Unfortunately, because the WSC measurements are in an anomalous section of the river, the calculated roughness (from WSC discharge measurements) is considerably higher than the reach-average value should be, even though the measured ice thickness is probably a fair indication of the average ice thickness in the reach. However, one available velocity profile (Fig. 9.8) indicates that the hydraulic radii in the
ice- and bed- affected flow are exactly one-half the mean flow depth for that one vertical. This suggests that the roughness of the bottom of the ice cover is the same as that of the bed, which ultimately leads to the same composite hydraulic roughness.

The question remains, however, as to the appropriate value of the composite roughness for a newly formed ice cover. This is addressed in the following section which describes one particular freeze-up event that has been documented in detail.

9.4. COMPOSITE ROUGHNESS - 1981/82 MEASUREMENTS

A unique combination of events produced near flood conditions at Peace River town in early January, 1982 during the formation of a stable ice cover. The resulting concern provided the necessary impetus to undertake a series of measurements to better understand the phenomenon. These included measurements of ice thickness, observations of the ice front progression, and a careful assessment of discharges. Summer surveys were also undertaken to evaluate the hydraulic characteristics.

9.4.1. Description of Freeze-Up Sequence

Ice conditions were first observed on December 18 when the ice front was located some 145 km downstream of Peace River and progressing upstream at a rate of about 0.16 m/s (Acres, 1984). A stable ice cover formed at Peace River on January 2, producing an increase in water level from about 312.0 m to a peak elevation of 314.8 m at a mean daily discharge of about 1860 m³/s and a mean daily air temperature of -31°C (Fig. 8.9).

With the mean daily air temperature dropping to -37°C and inflows from upstream briefly reduced to 1100 m³/s, the ice cover progressed upstream at a rate of about 0.36 m/s for the next five days. Assuming the stage increase of about 2 m at Peace River was typical for the reach just upstream of the gauge, approximately 360 m³/s of flow was being abstracted into storage as the cover progressed upstream from Peace River. Thus, the estimated mean daily discharge experienced at the gauge was reduced to about 1500 m³/s and the mean daily water levels dropped by about 0.3 m immediately following the passage of the ice cover on January 4.

The ice cover reached Dunvegan (102 km upstream of Peace River) early on January 6. With flows in the order of 1,500 m³/s, the stage increase due to the formation of the ice cover was measured to be approximately 3.8 m. It is difficult to determine the average stage increase over the entire reach because
the discharge (on which the stable ice thickness depends) varied considerably between 1100 m$^3$/s and 1800 m$^3$/s, averaging about 1400 m$^3$/s. However, a good approximation would be about 3 m of backwater, composed of both ice and water. With an average channel width of 450 m, about 400 m$^3$/s was being lost to channel storage over the four days it took the ice cover to reach Dunvegan.

On January 7, after the ice cover had progressed some distance upstream of Dunvegan, it was apparent that the sudden increase in discharge due to the fluctuating releases from the Bennett Dam produced massive consolidation of the ice cover. Initially, a 9 m high jam formed at a location about 20 km downstream of Dunvegan. This jam quickly failed and a surge of ice and water moved downstream (Næstdal, 1982), reaching Peace River at 22:30 hours (Fig. 9.9) and causing almost an immediate increase in stage of 3.54 m to a maximum water elevation of 318.15 m, some 3.4 m above the level of the previous stable ice cover. Within 2 hours of the peak, the stage dropped by 0.60 m and after about 36 hours the stage dropped a further 1.15 m to an elevation of 316.4 m where it remained constant for the rest of January. From an aerial inspection at a later date, it was apparent that effects from the consolidation extended to a point about 30 km downstream of Peace River although the thickened ice cover associated with the high discharge of the flood wave had dissipated within 8 km of the town.

On January 8, twelve hours after the peak at Peace River, the head of the cover was observed at km 880 (about 40 km downstream of the town), progressing upstream at 0.17 m/s (Acres, 1984). This rate was maintained at least until January 11, but between then and January 15 the upstream progression slowed to 0.04 m/s perhaps due to the warmer temperatures. On January 16 the cover reestablished its upstream progression at a rate of 0.18 m/s and the head moved past Dunvegan during that night. The stage increase associated with the cover formation was 4.7 m for a discharge of about 1770 m$^3$/s and a mean daily temperature of -25°C.

Assuming that a backwater of 4.7 m was typical of the average backwater in the reach between Peace River and Dunvegan, the ice front velocity of 0.17 m/s results in a loss of flow to channel storage of about 400 m$^3$/s for the period of January 8 to 11 and January 16 to 19. For the period of January 12 to 15, when the velocity was only 0.04 m/s, about 100 m$^3$/s of flow was being lost to channel storage. This interpretation appears to be consistent with the response of the gauge at Peace River (Fig. 9.9).

### 9.4.2. Ice Cover Characteristics

A high water profile, the prevailing water level profile, and ice thickness measurements, were obtained on January 13, 1982 (Figs. 9.10 and 9.11). The thickness measurements indicated a relatively consistent submerged thickness of between 3.8 m and 4.2 m, although in some locations the submerged portion of the cover was as thin as 2.3 m. The cover was formed primarily from frazil slush in which were imbedded solid broken ice floes originating from either broken border ice or frozen crests of the frazil pans (Fig. 9.12a). The broken border ice ranged between 0.5 and 1.0 m in thickness and the frozen crests were in the order of 0.3 m thick.

The maximum ice push up the bank along the study reach was between 0.9 and 1.5 m above the January 13 water level, which is in the same order as the maximum water level associated with the ice run. The perceived average top of ice on the day of the survey was generally between 0.2 to 0.6 m above the water level and where shear lines were evident, ice had pushed up at least 1.6 m above the water level (Fig. 9.12b). The surface of the cover was composed of hummocky, drained, frozen porous frazil floes which varied in height up to 0.7 m above the water level. The mean relief of 0.4 m corresponded with the mean of ice level estimates at other locations and hence it was typical of the ice cover for the entire reach (Fig. 9.12a).

The winter ice thickness measurements were augmented by measurements of shear walls at breakup after the passage of the ice front. Many of the exposed shear walls were still intact and, because they cantilevered out above the bed, their thickness could be easily measured. These thicknesses ranged from 2 m to 5 m. Most of the shear walls were about 4 m thick and the thinner ones were probably grounded out. The reliability of these measurements is not as great as for the winter measurements but they do substantiate and augment the latter.

From the measurements and observations of shear lines and relative levels of the assorted ice formations, it was evident that the distribution of the ice was extremely variable. Many of the back channels which carry flow during the summer were not contributing to the conveyance of the flow. Similarly many of the shallow areas in the lee of islands and on the inside of bends also probably did not contribute. This made the interpretation of the effective ice thickness and conveyance characteristics of the channel extremely subjective.
Fig. 9.10. Longitudinal profile in the study reach.

LEGEND
- Peak water level, Jan 7, 1982
- Surveyed water level, Jan 13, 1982
- Surveyed water level, July, 1982
- Measured shear wall thickness
- Bottom of consolidated ice cover
- Ice
- Water
- Bedrock

Q = 275 m³/s
Q = 1000 m³/s

Fig. 9.11. Typical cross sections and measured ice thickness in the study reach.
9.4.3. Hydraulic Characteristics

The discharge increase and force of the moving ice associated with the surge lifted and disintegrated the newly formed ice cover, moved it downstream, and consolidated it into a porous mass of ice with a submerged aggregate thickness of about 4 m. This increased thickness, along with its associated roughness was the main reason for the extremely high stage experienced at the town. Figure 9.10 illustrates the measured ice jam profile, the highwater profile, and the ice thickness measurements. The water level profile of the stable ice cover, the highwater marks, and the adopted bottom of the ice cover are parallel to the open water profile, with a slope of 0.29 m/km. This suggests that quasi-uniform (variable longitudinally but without a trend) flow prevailed for all three measured conditions. The average hydraulic characteristics for two open water conditions and the surveyed ice condition in mid-January are summarized in Table 9.1.

<table>
<thead>
<tr>
<th>Characteristics</th>
<th>Open Water</th>
<th>Ice Cover (January, 1982)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Discharge, Q (m³/s)</td>
<td>1100</td>
<td>2750</td>
</tr>
<tr>
<td>Top width, W (m)</td>
<td>510</td>
<td>540</td>
</tr>
<tr>
<td>Flow area, A (m²)</td>
<td>1250</td>
<td>1800</td>
</tr>
<tr>
<td>Mean depth, h (m)</td>
<td>2.45</td>
<td>3.33</td>
</tr>
<tr>
<td>Hydraulic radius, R (m)</td>
<td>2.45</td>
<td>3.33</td>
</tr>
<tr>
<td>Mean velocity, (m/s)</td>
<td>0.88</td>
<td>1.4</td>
</tr>
<tr>
<td>Ice thickness, t (m)</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Manning roughness n₀</td>
<td>0.035</td>
<td>0.025</td>
</tr>
</tbody>
</table>

The analysis summarized in Table 9.1 implies that a composite Manning roughness coefficient of 0.027 is appropriate for a post-freeze-up ice cover near Peace River. This suggests that the bed roughness and ice roughness are approximately equal. There may be a trend of increasing roughness coefficient with decreasing flow depth under the ice cover, in the same manner as this coefficient increases with decreasing flow depth for open water conditions. However, there is insufficient data to confirm this. For the purpose of design, a composite roughness coefficient of 0.027 is probably appropriate because most severe freeze-up ice levels exhibit flow depths in the order of 3 to 4 m.
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It might seem possible to simply extrapolate the data to the peak stage of the event, and calculate both the peak discharge and the dimensionless coefficient of internal friction of the consolidated cover. Unfortunately, the flow was highly unsteady during the consolidation and a much more complex unsteady flow analysis would be required to define the relationship between stage, discharge, the force on the ice cover, and the evolution of the thicker cover. It is possible that the maximum discharge and ice thickness occurred prior to the peak stage due to the characteristics of the flood wave associated with the consolidation (M. Ferrick, pers. commun.). Thus a steady, uniform flow analysis at the peak stage would give misleading estimates of discharge. However, this crude analysis suggests a peak discharge in the order of 2500 m³/s, at a peak water depth of 8.7 m for a total ice thickness of 4.3 m and a total flow depth of 4.7 m.

9.5. COEFFICIENT OF INTERNAL FRICTION - ANNUAL FREEZE-UP LEVELS

Prior to regulation, when freeze-up discharges were relatively low, juxtaposition could have been the dominant process in some years. However, with an increase in the discharge during freeze-up in the years following regulation, observations indicate that the mode of freeze-up is consistent from year to year and showing seems to dominate. Furthermore, although the process of ice cover formation is unsteady, the resulting ice cover in the reach of interest is the result of a quasi-steady process for a duration of at most one day during which a uniform ice cover forms throughout the reach. Also, the 1981/82 measurements suggest that the stage measured at the WSC gauge at Peace River town is representative of the stage within the entire reach of interest. This allows the use of reach-averaged channel characteristics in defining the hydraulic characteristics of the flow under the ice cover.

Table 9.2 summarizes the post-regulation freeze-up levels, along with the discharges and the reach-average hydraulic characteristics for each event. As noted in the table, the freeze-up elevation was taken as the maximum nominal freeze-up level to account for the shoaling phenomenon. However, the small short duration peaks above this level produced from local surging were discounted in the analysis. The discharge was estimated by routing flows from upstream or by applying the open water rating curve to the measured water level at the WSC gauge just prior to the development of any backwater from ice formation downstream. The total water depth was easily determined from Fig. 9.5 and the flow depth under the ice cover was computed using reach-average values of width.

### Table 9.2 Summary of post regulation freeze-up characteristics

<table>
<thead>
<tr>
<th>Year</th>
<th>Freeze-up Date</th>
<th>Freeze-up Elev (m)</th>
<th>Discharge m³/s</th>
<th>Total Depth (m)</th>
<th>Flow Depth (m)</th>
<th>Ice Thickness (m)</th>
<th>Measured Ice Thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1979</td>
<td>Dec 15/79</td>
<td>311.8</td>
<td>700</td>
<td>2.0</td>
<td>2.1</td>
<td>1.0</td>
<td>0.8 (7)</td>
</tr>
<tr>
<td>1980</td>
<td>Dec 15/80</td>
<td>334.0</td>
<td>800</td>
<td>2.7</td>
<td>2.3</td>
<td>1.2</td>
<td>1.6 (20)</td>
</tr>
<tr>
<td>1981</td>
<td>Dec 15/81</td>
<td>319.6</td>
<td>1550</td>
<td>5.4</td>
<td>3.4</td>
<td>2.1</td>
<td>1.9 (19)</td>
</tr>
<tr>
<td>1982</td>
<td>Dec 8/82</td>
<td>334.0</td>
<td>1400</td>
<td>4.9</td>
<td>3.2</td>
<td>1.4</td>
<td>2.2 (26)</td>
</tr>
<tr>
<td>1983</td>
<td>Dec 27/83</td>
<td>316.4</td>
<td>1100</td>
<td>5.2</td>
<td>3.8</td>
<td>2.4</td>
<td>2.0 (29)</td>
</tr>
<tr>
<td>1984</td>
<td>Jan 27/84</td>
<td>333.0</td>
<td>1900</td>
<td>3.9</td>
<td>3.1</td>
<td>1.5</td>
<td>1.5 (28)</td>
</tr>
<tr>
<td>1985</td>
<td>Jan 13/85</td>
<td>316.4</td>
<td>2000</td>
<td>5.3</td>
<td>4.0</td>
<td>2.0</td>
<td>1.3 (24)</td>
</tr>
<tr>
<td>1986</td>
<td>Jan 24/86</td>
<td>312.5</td>
<td>900</td>
<td>5.3</td>
<td>2.1</td>
<td>1.3</td>
<td>1.2 (19)</td>
</tr>
<tr>
<td>1987</td>
<td>Dec 15/87</td>
<td>314.0</td>
<td>1100</td>
<td>4.9</td>
<td>3.8</td>
<td>2.3</td>
<td>2.3 (29)</td>
</tr>
<tr>
<td>1988</td>
<td>Jan 2/88</td>
<td>314.2</td>
<td>1400</td>
<td>5.0</td>
<td>3.2</td>
<td>2.0</td>
<td>2.1 (29)</td>
</tr>
<tr>
<td>1989</td>
<td>Jan 5/89</td>
<td>315.2</td>
<td>1500</td>
<td>6.0</td>
<td>3.4</td>
<td>2.8</td>
<td>2.4 (16)</td>
</tr>
<tr>
<td>1990</td>
<td>Jan 18/90</td>
<td>316.4</td>
<td>1600</td>
<td>5.3</td>
<td>3.5</td>
<td>3.0</td>
<td>1.5 (20)</td>
</tr>
<tr>
<td>1991</td>
<td>Dec 21/91</td>
<td>315.9</td>
<td>1700</td>
<td>6.7</td>
<td>3.6</td>
<td>3.4</td>
<td>3.1 (20)</td>
</tr>
</tbody>
</table>

1 Discharge estimated from discharge upstream or from water levels at gauge one or two days prior to freeze-up.
2 Total depth determined from Fig. 9.5.
3 Flow depths computed using a composite roughness of 0.027 and the reach-average width and slope.
4 Ice thickness computed as distance between total water depth and flow depth divided by the specific gravity of ice.
5 Reservoir run-off included the number of days after freeze-up when the first discharge measurement was made.

and slope and a composite roughness of 0.027. This then allowed for a computation of the reach-average freeze-up ice thickness which can be compared to measured ice thickness made by WSC at only one section, at a time some 7 to 29 days following freeze-up. Comparison of the computed ice thickness with that measured suggests that the computed thicknesses are not unrealistic (Table 9.2). If anything, they may be on the high side. That is the composite roughness has been assumed to be too low. However, given the propensity for the frazil accumulation to thin out with time, the computed thickness seems to be in the appropriate range.

With known ice thickness, flow depth, width, channel slope, and the
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hydrologic radius associated with the ice cover, it is possible to compute the dimensionless coefficient of internal stability, $\mu$, for each freeze-up event. To do this, Eq. 4.36, Chapter 4, may be used, assuming zero cohesion. The results are shown in Fig. 9.13. The stage increase associated with the formation of the ice covers varies from less than 1 m for discharges less than 1000 m$^3$/s, to 4 m for discharges in the order of 2000 m$^3$/s. With an assumed composite Manning roughness coefficient of 0.027, the observed freeze-up stages can be reproduced by choosing different values of $\mu$ in the range of 0.8 to 2.0. Figure 9.13 shows that $\mu$ is much larger for freeze-up events associated with discharges under 1200 m$^3$/s. The low flow probably allows downward freezing in the pack to increase the internal strength, thus producing a higher apparent value of $\mu$. When the discharge is greater than 1500 m$^3$/s, the shoving process

![Fig. 9.13. Freeze-up rating curve and evaluation of the dimensionless coefficient of internal friction.](image)

seems to dominate and $\mu$ has to be in the range of 0.8 to 1.5 in order to predict the water levels associated with these conditions (see also Chapter 4, Section 4.3.7). It is interesting that this variation of about 100% between the minimum and maximum values of $\mu$ produces a difference of about 1 m in the ultimate stage of a freeze-up ice jam.

9.6. DESIGN IMPLICATIONS

The Peace River in the vicinity of Peace River town is a large river regulated for hydropower generation. The river is relatively steep and, at high discharges,

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shoving dominates the ice cover formation process. This generates a relatively thick cover but with a significant depth of flow under it. For these conditions, at this location and other locations like it, the nominal freeze-up stage resulting from the formation of a frazil jam is often the highest flood stage of the year. Knowing the discharge regime, this flood stage can be calculated using a composite roughness of 0.027 and a dimensionless coefficient of internal friction in the range of 0.8 to 1.5. For conservative applications for the events with higher discharges (for which the highest stages will result) it is recommended that the range of $\mu$ be further reduced to 0.8 to 1.0.

Given the high probability for an equilibrium section of the jam to exist anywhere within the reach of interest, a profile can be computed using the model HEC-2 by defining the composite Manning coefficient and the appropriate ice jam thickness. A simpler technique which is certainly adequate and as accurate, providing the reach is relatively uniform, is the uniform flow approximation, using any acceptable resistance equation.

There are some questions about the impact of the short-duration surges which may thicken the cover beyond that achieved by the effect of the quasi steady flows and which also produce temporary stage increases above the design values. In some cases, such as in 1981/82, these may be significant. Fortunately such an event is related to drastic fluctuations in upstream reservoir releases and in any rational operation would not be tolerated. For the most part the surges produce temporary increases in stage of about 0.5 to 0.75 m. This phenomenon may be taken into account by nominally increasing the expected mean daily discharge by 25% and then making the computation to determine the upper limit of the stage of the frazil jam under the assumption of steady flows.

9.7. CONCLUSIONS

This case study has illustrated one ice jam problem in Northwestern Canada and the extent of the data set typical of the majority of these situations. Examples of some of the techniques used to analyze these data to achieve a particular end product - in this case an estimate of the expected flood levels resulting from the development of an ice cover due to the formation of frazil jams - are included.

The general freeze-up process has been discussed. One freeze-up event which was more closely documented and analyzed in order to determine the composite roughness of the ice covered channel has also been described. In
addition, the discharges and water levels associated with freeze-up for each year were analyzed to determine the variability of the freeze-up ice thickness and the coefficient of internal friction of the frazil accumulation.

For this particular case study it was found that the bed roughness is approximately the same as the roughness of the underside of the ice cover for flow depths which are typical of most freeze-up situations. The resulting composite roughness is 0.027 and the coefficient of internal friction varies between 0.8 and 1.5 for the events when the cover is obviously formed by shoving. For more conservative design criteria it is recommended that the coefficient of internal friction be taken as 0.8 and that the quasi-steady discharge at freeze-up be increased by 25% to account for short term water level increases due to the surges associated with the dynamic nature of the ice cover formation.

APPENDICES

Spyros Beltaos

APPENDIX A. STRENGTH CHARACTERISTICS OF ICE JAMS

Consider first the loading configuration used by Uzuner and Kennedy (1974) and Cheng and Tatnall (1979) to determine the compressive strength of a floating accumulation of ice blocks. A longitudinal force was applied to one end of the accumulation which floated in stagnant water while being confined laterally. The longitudinal force was applied by a moving vertical plate and it was intended that shear on the sides of the accumulation would be minimal. Assuming completely frictionless sides, the x-, y-, z- stresses would be principal ones and a plane-strain condition would prevail, i.e., \( s_x, s_y, \) and \( s_z \) would be independent of \( x \) (transverse coordinate). When the accumulation is undergoing failure, the Mohr-Coulomb criterion is assumed to apply everywhere (Uzuner and Kennedy 1974). Assuming that \( s_x \) (= vertical stress) and \( s_y \) (= longitudinal stress) are the least and largest principal stresses, respectively, the Mohr diagram would be as shown in Fig. A.1 which implies (Uzuner and Kennedy 1974):

\[
x = \frac{2c_s \cos \phi}{1 + \sin \phi} \frac{1 + \sin \phi}{1 - \sin \phi} \quad (A.1)
\]

Taking the vertical averages of both sides, gives

\[
\sigma_x = \frac{2c_s \cos \phi}{1 + \sin \phi} \quad (A.2)
\]

The value of \( \sigma_x \) may be obtained by vertical integration of Eq. 4.19b in Chapter 4, after omitting the shear terms (=0 in this case):

\[
x = \begin{cases} 
\gamma(1 - s_x)(1 - p_x)(x - y_{h_2}) & \text{for } y_{h_2} \leq y \leq y_x \\
 \gamma(1 - s_x)(1 - p_x) \gamma x - \gamma p_x (1 - p_x)(x - y) & \text{for } y_x \leq y \leq y_r 
\end{cases} 
\quad (A.3)
\]