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Application of Numerical Modelling Towards Prediction of River Ice Formation on the Upper Nelson River

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In recent years, a concentrated effort has been applied in the Province of Manitoba towards the practice of river ice forecasting during both freeze-up and break-up. This research focuses on freeze-up forecasting on a reach of the Upper Nelson River upstream of the Jenpeg Generating Station, where hanging dams, intake blockages, and other frazil ice constrictions have caused significant generation losses in the past three decades. Prediction of the river ice processes is made using the CRISSP2D finite element model. This paper outlines the current status of the project, including: model development, mesh generation, and the calibration of hydrodynamic, linear heat exchange, and static ice parameters used by the model.

1. Introduction

The prevention or mitigation of ice-induced problems during freeze-up often takes precedence in the winter operation of hydroelectric generating stations and hydraulic control structures in cold regions. In Manitoba, frazil ice has been found to significantly restrict the winter operation of northern hydroelectric generating stations through channel constrictions (Zbigniewicz, 1997) or anchor ice formation (Malenchak *et al.*, 2006). As a result, concentrated effort has been applied to the practice of river ice forecasting on both municipal and provincial levels in order to better understand ice phenomena and minimize any associated financial and social risks.

The focus of this research is on the application of numerical modeling in predicting river ice formation upstream of the Jenpeg Generating Station and Control Structure. Located along the Upper Nelson River, Jenpeg provides primary control of Lake Winnipeg outflow and allows for its use as a natural storage reservoir. Regulation of the lake is particularly important during winter months, when energy demand is the highest and the flow is the lowest. The largest operating risk to the station is frazil ice; the station intakes have been either partially or fully blocked by frazil numerous times in the past thirty years.

1.1 Lake Winnipeg Regulation

The regulation of Lake Winnipeg was achieved with the completion of the Lake Winnipeg Regulation (LWR) Project in 1976. The primary goal of the project was to establish Lake Winnipeg as a long-term hydroelectric storage reservoir and facilitate the development of the hydroelectric potential of the Lower Nelson River. In addition to the Jenpeg Generating Station and Control Structure, the project included the construction of several channel excavations and improvements that were aimed at increasing total outflow from the lake during ice-on conditions. As outlined in Figure 1, this included the 2-Mile, the 8-Mile, and the Ominawin Bypass Channels, as well as minor dredging works at various constrictions in the reach.

1.2 Ice Stabilization Program

The LWR Project also includes an operation strategy for Jenpeg during freeze-up aimed at monitoring and controlling the ice formation processes in the channels upstream of the station. Based on daily observations of flowrate, water temperature, ice conditions, and a short-term weather forecast, a flow cutback is initiated before any significant frazil production takes place. In doing so, the formation of a stable static ice cover is encouraged and any long term problems, such as the formation of hanging dams or intake blockages at the station, are avoided. A maximum flow of 1650 cms is necessary to achieve these goals, representing an average reduction of 1000 cms (Tuthill, 1999).

2. Study Area

Jenpeg is located roughly 130 km downstream of Lake Winnipeg along the Nelson River West Channel, as shown in Figure 1. The study area for this project encompasses an area of 315 km^2, spanning from the Jenpeg forebay to Playgreen and Kiskittogisu Lakes. The hydrology in this region is complex, consisting of shallow lakes connected by several rock-controlled channels. Flow leaving the two upstream lakes is split between the Ominawin (50%), Metchanais (30%), and Kisipatchewuk (20%) Channels before remerging to form the Nelson River West Channel. From there, the flow passes through Manitou and Saskatchewan Rapids before reaching Jenpeg.

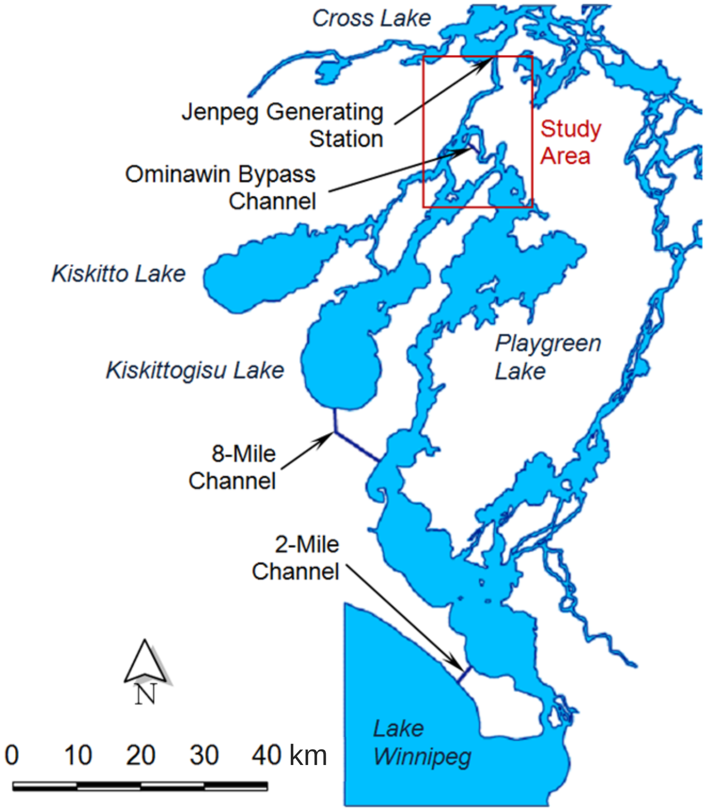


Figure 1. Overview of the Lake Winnipeg Regulation Project.

The conventional ice regime consists primarily of static ice formation on all of the larger lakes and bays. Sections of open water in high velocity areas stay open late into freeze-up, forming either skim or frazil ice. The type of ice that is formed is dependent largely on the flow passing the station and may vary from one year to the next. An ice front begins at the ice boom located in the forebay and its upstream progression due to the juxtaposition of surface ice floes closes the remaining open water sections. Typically, this process requires several weeks to complete; however, the large majority of ice generation takes place in the two to three days directly following the flow cutback.

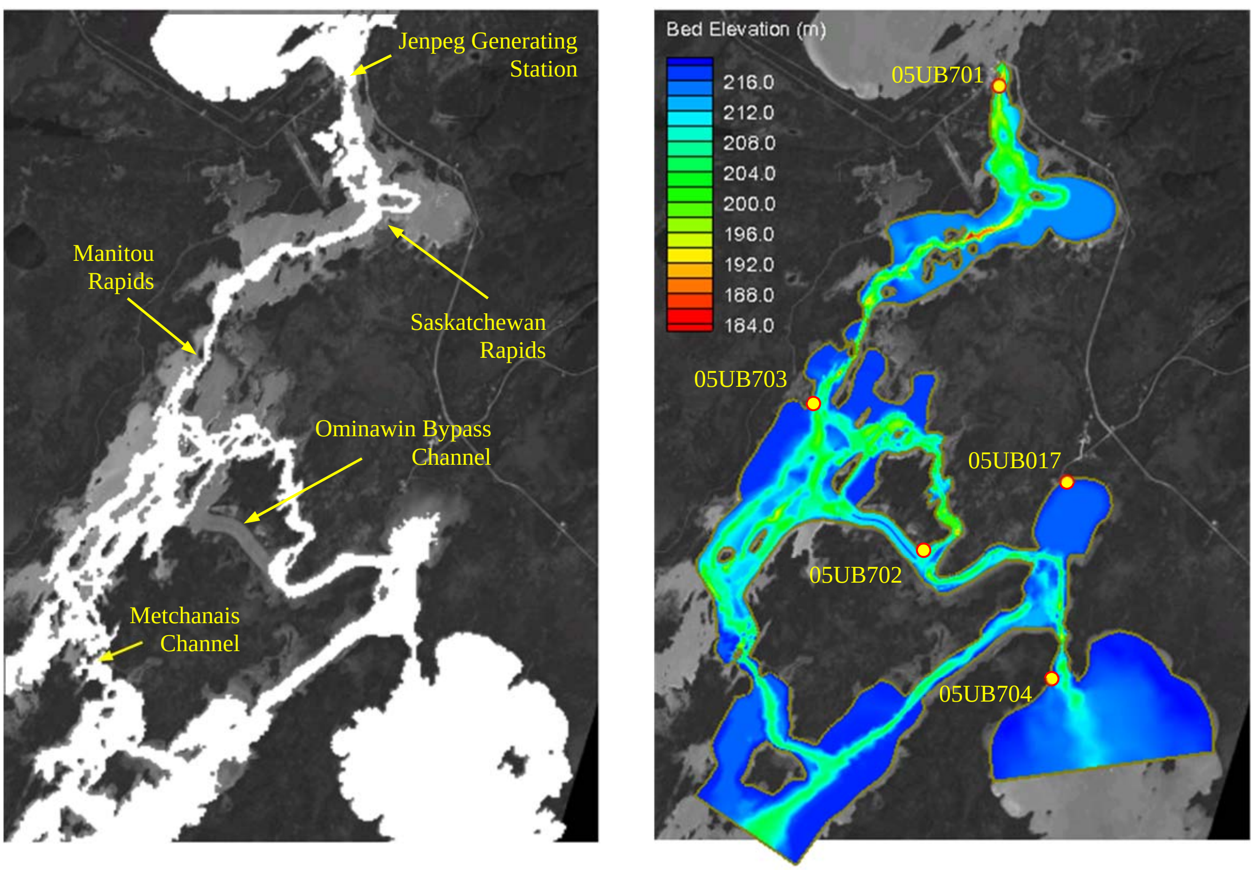
3. Model Development

The CRISSP2D finite element model was utilized for the purposes of this study. Ice generation is calculated through the coupling of three modules, simulating the hydrodynamics, thermodynamics, and ice dynamics of the channel. The hydrodynamic module provides an estimation of water depth, water velocity, and other hydraulic properties at each finite element node. The thermodynamic module calculates heat transfer at the water/air boundary and uses the results to predict border ice or skim ice formation based on a critical value criteria developed by Michel *et al.* (1982) and Matousek (1984). Lastly, the dynamic ice module enables simulation of frazil ice using a concept of thermal equilibrium, as well as mass exchange processes that take place along the channel bed or water surface (Liu et al., 2005). This includes the formation of anchor ice, frazil accretion to the border ice edge, or ice boom interactions.

3.1 Mesh Generation

CRISSP2D requires that the model domain be discretized into a network of triangular finite elements. Such a mesh allows for a high degree of flexibility in defining element size, shape, or density where complex geometry or other channel irregularities warrant a higher resolution. For this project, two mesh densities were analyzed for both computation time and accuracy. The first, a 6,500 node mesh, showed to be computationally efficient but was highly unstable and inaccurate. The largest errors were realized in high velocity channels with low cross-sectional node resolution (6 nodes). To mitigate this, the entire domain was refined such that all main channels were defined by a minimum of 9 cross-sectional nodes. This produced a 13,250 node mesh and decreased model error from 40% to 1%. The final mesh resolution varied from 10 m in narrow or complex channels to 500 m in shallow lakes and large bays.

Bathymetric information for the model was provided by a series of 2-dimensional cross-sections taken pre-construction. This data was mapped and converted into a 3-dimensional georeferenced scatter set, and was interpolated onto the mesh using an inverse distance weighted scheme. In areas where the channel thalweg, shown in Figure 2a, was not properly captured, one or more proxies of the nearest cross-section were added and the mesh was re-generated. In total, 143 proxy cross-sections were used in conjunction with the original 209. The final mesh along with the data sources to which the model was calibrated is shown in Figure 2b.



a) Location of channel thalweg b) Model bathymetry and data sources

Figure 2. Overview of the CRISSP2D model domain.

3.2 Boundary Condition

The model is forced at each of its three open boundaries: one water elevation boundary downstream at the Jenpeg Generating Station, and two flow boundaries upstream at Playgreen and Kiskittogisu Lakes. Station 05UB701 is used directly to estimate the downstream elevation boundary and the estimated flow at Jenpeg is split between each of the two upstream boundaries. Based on physical measurements, this split was found to be roughly 60/40 between Playgreen Lake and Kiskittogisu Lake, respectively.

For thermodynamic and dynamic ice simulations, incoming water temperature and ice concentration are also required. Water temperatures from each of the five gauges are used to provide an approximation, which is then applied at both upstream boundaries. It is assumed that once the water temperatures have stabilized at freezing, water entering the model is at 0.01°C. In addition, given that the upstream lakes freeze thermally, it can be assumed that the incoming ice concentration is negligible. Both of these assumptions allow for rather simple and consistent boundary definitions for a majority of ice simulations conducted.

4. Hydrodynamic Simulations

Model calibration was performed on the 2008 open water season. This timeframe was chosen due to the highly dynamic flow and elevation changes observed at the station. Measured flow at the station varied from a low of 1640 cms to a high of 3800 cms, whereas the forebay elevation cycled between 217.2 m and 213.8 m. Subsequently, the open water seasons for years 2001, 2002, 2004-2007, and 2009 were used as verification. This included both low flow (2004) and high flow (2005) years, and modeled the entire licensed operating range at the station, including minimal flows typically encountered during the flow cutback.

To aid in model calibration, the mesh was segregated into 24 unique reaches defining areas of different gradient, bed material, bathymetric features, or other hydraulically significant characteristics. The Manning bed roughness coefficient was then adjusted for each reach until an optimal match of simulated to observed water elevations was achieved. Two areas required a subsequent minor adjustment to bed elevation to achieve a good fit. These included a 500 m long reach adjacent and downstream of station 05UB703, and a 2000 m long reach directly upstream of the Upper Ominawin Channel. In both situations, the bed elevation was raised by approximately 1 meter.

Typical Manning bed roughness in the model was determined to lie in the range of 0.022-0.040 for main channel sections and 0.050 for flooded areas. A low roughness of 0.015 was used in the immediate forebay area. An artificially high roughness of 0.123 was applied to the Metchanais channel to account for constriction head losses not captured by the model due to the exclusion of several small islands in the narrow channel.

The combined results from the final calibration and verification simulations are presented in Figure 3 for each gauge. Results from gauge 05UB701 have been omitted given its close proximity to the elevation boundary and excellent performance ($R^2 = 0.9997$). The model results exhibit increased variability progressing upstream, primarily due to a slight delay in model response to flow changes. The largest degree of variability occurs at the furthest upstream gauge located on Playgreen Lake, where the influence of Lake Winnipeg wind effects are significant.

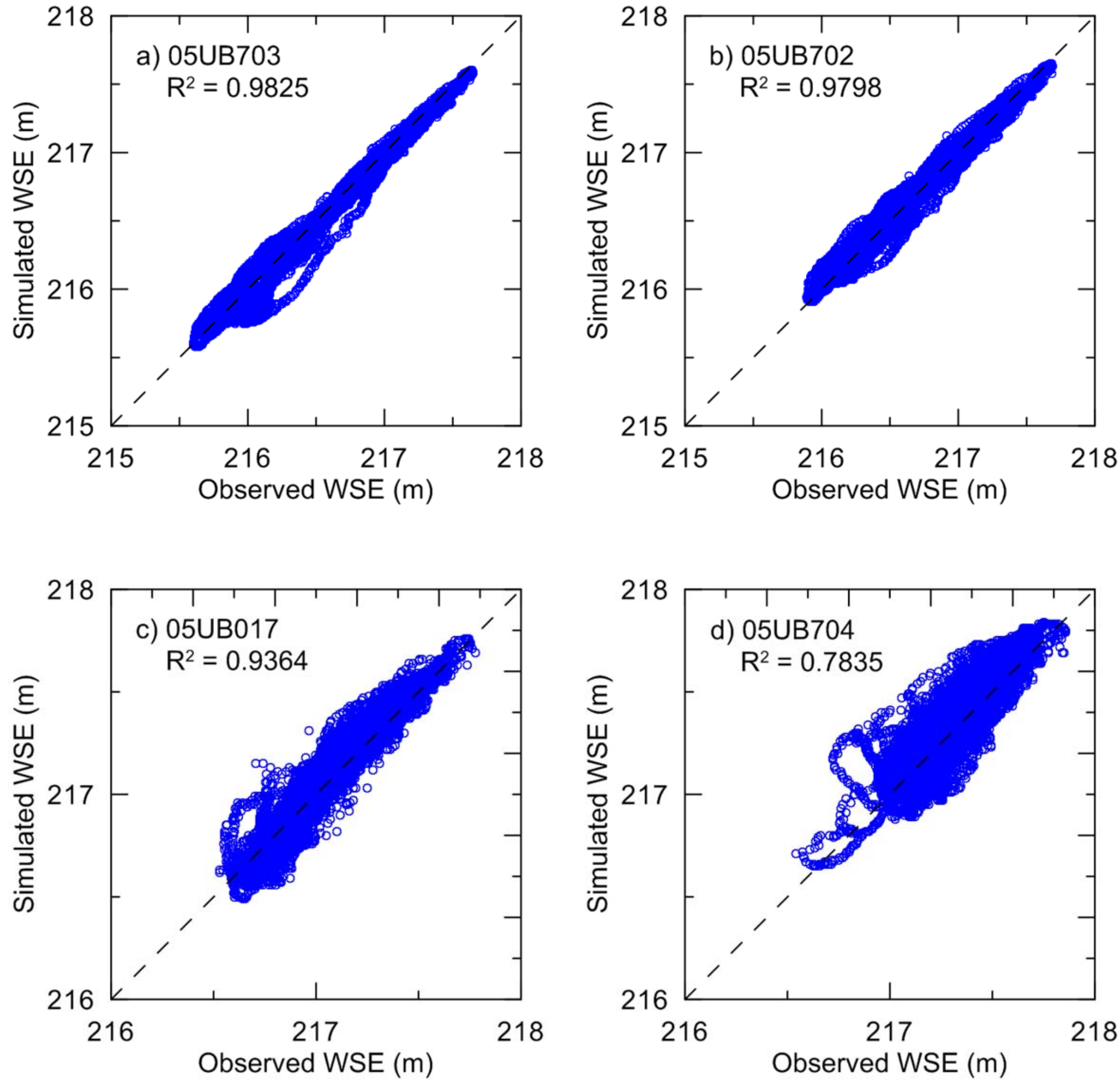


Figure 3. Performance plots of all hydrodynamic simulations.

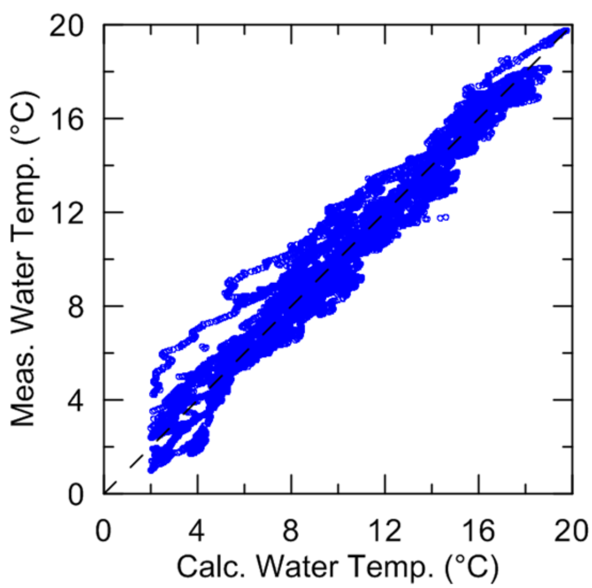
5. Surface Heat Exchange

The onset of freezing on rivers and lakes is primarily dependent on the gradual heat loss that takes place at the water surface. Within CRISSP, this process can be modeled using either a linear heat transfer approach or a full energy budget. For the purposes of this study, a lack of full meteorological data required the use of linear heat transfer, which utilizes a bulk heat transfer coefficient term to define water temperature changes based solely on the air-water temperature differential in each time step. This method does not directly consider the influence of wind, solar radiation, precipitation, or other important parameters affecting water temperature.

Calibration of this parameter was performed independent of CRISSP using an average water temperature and a heat transfer function, defined as follows:

$$T_{w,t} = \frac{C_0\left(T_{a,t-1} - T_{w,t-1}\right)}{C_{p\_w} \cdot H_{t-1} \cdot \rho} \cdot \Delta t \quad [1]$$

The average water temperature at each time step ($T_{w,t}$) was calculated based on the air ($T_{a,t-1}$) and water ($T_{w,t-1}$) differential in the previous time step. Other parameters include: the linear heat transfer coefficient (C_0), the specific heat of water (C_{p_w}), average water depth (H_{t-1}), water density (ρ), and the time step (Δt). Calibration of the linear heat transfer coefficient was carried out using each of the cooling periods in years 2001 to 2010. The same value was applied to each year and was simultaneously optimized using linear regression to yield a global value of 22.38 W/m^2°C, which is in agreement with the typical range of 15 to 25 W/m^2°C. A summary of the complete calibration results, as wells as those for each year are presented in Figure 4.



Year	Slope	R^2
2001	0.9805	0.977
2002	1.0489	0.974
2003	0.9680	0.989
2004	1.0473	0.622
2005	1.0038	0.983
2006	1.0029	0.947
2007	1.0080	0.939
2008	0.9462	0.979
2009	0.9834	0.970
2010	0.9313	0.944
Total	0.9896	0.958
1:1	1	0.957

Figure 4. Summary of the linear heat transfer calibration.

With the exception of 2004, excellent fit was achieved for each of the calibration years. Model performance is best for colder temperatures, whereas some degree of underestimation was noted at higher temperatures. The highest source of error was due to a shift in temperature change from cooling to warming, where a slight lag was observed to cause divergence of the result from measured values. This was the primary cause of error for the 2004 simulation, where four large temperature fluctuations occurred. Overall, the results exhibit a proportional trend with a best fit slope of 0.9896 and an R^2 of 0.958. Relative to a 1:1 slope, the R^2 reduces slightly to 0.957.

6. Static Ice Simulations

The simulation of static ice in CRISSP is primarily dependent on two parameters: one defining a critical velocity above which surface ice will not form (*vcrskm*), and the other a surface water temperature required for surface ice initiation (*tc*). Previous studies by Michel *et al.* (1982), Matousek (1984), Santeford (1990), and others have directly analyzed border ice growth in relation to surface velocity. It was found that thermal growth occurs only where water velocities do not exceed 0.4 to 0.5 m/s; values used in Manitoba have ranged as high as 0.7 m/s. Using this range, the rate and extent of the border ice edge can be predicted having knowledge of the surface velocity distribution.

The onset of freezing is largely dependent on the surface water temperature, which is estimated in CRISSP using the following relationship developed by Matousek (1984):

$$T_{surf} = T_{mean} + \frac{q_0}{1130 \cdot v + b \cdot w} \quad [2]$$

where the surface water temperature (T_{surf}) is dependent on the depth averaged water temperature (T_{mean}), the net heat flux at the water/air boundary (q_0), the local depth averaged water velocity (v), and wind direction (b) and velocity (w) parameters. Supercooling of the water surface is required for ice formation. Matousek (1984) suggests a value between 0°C and -1.1°C is necessary for skim ice formation, and any additional cooling causes sheet ice formation over flowing water. By definition, surface water temperature will vary with water velocity. Therefore, in finite element models where a two-dimensional water velocity distribution is known, Equation 2 can be discretized over the model domain and can be used to better predict the extent of border ice progression.

Calibration of the static ice parameters *vcrskm* and *tc* was conducted using the 2010 freeze-up period, lasting from November 17, 2010 to November 22, 2010. This time-frame provided a good basis for testing the static ice module due to the very clearly defined border ice edge observed prior to the flow cutback. Assessment of model performance was made based on its accuracy in predicting the timing and extent of static border ice progression. The limiting velocity for surface ice formation (*vcrskm*) was found to be 0.5 m/s, and the initiation surface water temperature (*tc*) was found to be -0.75°C. In both cases, the calibrated values are within the expected range.

Verification of the static ice module was performed using the 2007 freeze-up period, lasting from November 13, 2007 to November 21, 2007. Freeze-up consisted primarily of intermittent border ice growth during the first week, followed by skim ice generation. Once again, a clearly defined border ice edge was observed each day and provided a spatial and temporal reference to which to compare model results. A sample of this assessment is presented in Figures 5 to 9, in which the model output is compared to photographs taken on site.

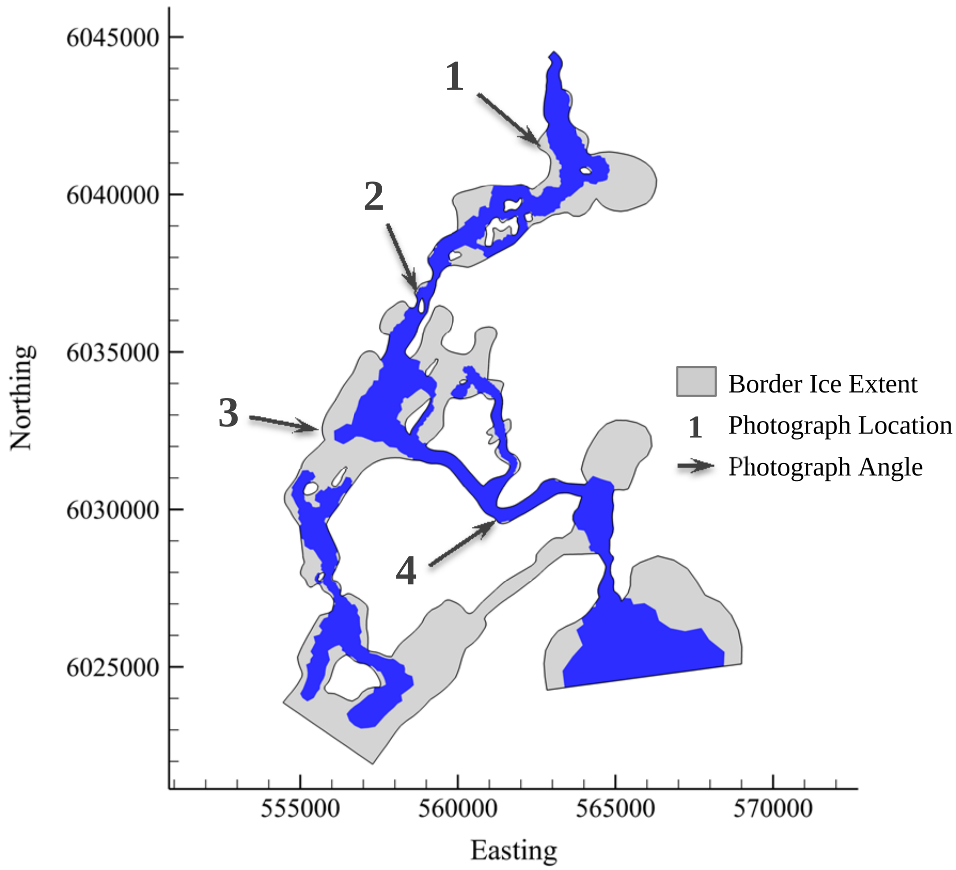


Figure 5. Extent of border ice progression, November 17, 2007

The model output for November 17, 2007 is presented in Figure 5. The four points outline areas on interest in regards to border ice formation. The first is located at Saskatchewan Rapids, near a flooded bay. Corresponding to Figure 6, ice is visible within the bay whereas the channel on all sides of the island remains open. Small patches of border ice are also visible on the east bank of the channel. This is accurately captured in the model, including the uneven open water widths on the channels north and south of the island.

The second point is located at Manitou Rapids and corresponds to Figure 7. Here, much of the channel remains free of ice late into freeze-up. Several isolated bays along the channel, as well as the bays east of Manitou Rapids freeze over early in the season. The third point, corresponding to Figure 8, looks at the Ominawin Bypass Channel. Two well-defined open water sections are visible at its outlet: a wider main channel on the west of the island, and a narrower secondary channel on the east. In addition, border ice is observed along the west bank of the island. Each of these observations are closely matched by the model output.

The last point looks at the Upper Ominawin Channel where high surface velocities inhibit any border ice growth. This is also evident on Figure 9, where minimal border ice formation has taken place. Further upstream, the large bay, which typically is the first to form ice in the region, has completely frozen over at this time.

Figure 6. Ice formation at Saskatchewan Rapids (Point 1), November 17, 2007

Figure 7. Ice formation at Manitou Rapids (Point 2), November 17, 2007

Figure 8. Ice formation at the Ominawin Bypass Channel (Point 3), November 17, 2007

Figure 9. Ice formation on the Upper Ominawin Channel (Point 4), November 17, 2007

8. Conclusion

Application of 2-dimensional modeling of river ice formation has become a major component of the planning, design, and operation of riverine infrastructure in northern climates. This research applies the CRISSP2D model on a portion of the Upper Nelson River upstream of the Jenpeg Generating Station in an effort to assess its ability in predicting both local river ice phenomena and the overall ice regime. Hydrodynamic calibration and verification was conducted on the open water seasons of years 2001, 2002, and 2004-2009 and showed excellent model performance at all flow conditions. Prediction of water temperature change was performed through a linear heat transfer approach where, through calibration, a global heat transfer coefficient of 22.38 $W/m^2°C$ was determined. Lastly, the static ice module within CRISSP2D was calibrated and verified. The two governing parameters were both within their expected range: the critical velocity above which surface ice does not form was found to be 0.5 m/s, and a surface water temperature of -0.75°C was found to be necessary for surface ice initiation. Both the timing and extent of border ice growth were accurately predicted for all years analyzed. Further model development includes its application under dynamic ice conditions and prediction of ice front progression at the ice boom.

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