ON THE SEDIMENT TRANSPORT CAPACITY OF RIVERS DURING ICE BREAKUP

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ABSTRACT

Rivers transport large quantities of sediment during ice breakup. This transport causes physical, chemical and biological changes to occur in the water, or on the river floodplains, banks and bed. There are several breakup processes that initiate sediment motion or enhance transport. In this paper we quantify the bed load and suspended sediment transport capacities of an ice-covered river with the shear stress on the bed and a composite shear stress, respectively, obtained from a model that considers coupled flow and ice motion. The abrupt motion of a river ice cover decreases the resistance to the flow, causing surges to develop that significantly increase the shear stresses and the sediment transport capacities. A case study of breakup on the Connecticut River is generalized by varying model input parameters that may be uncertain, including the relative roughness of the ice and the bed, the ice velocity and acceleration through time, the flow velocity prior to ice motion, and the energy gradient of the flow. The results presented include the surging flow velocities, dimensionless ice/flow velocity ratios, and dimensionless shear stresses for the ice zone, bed zone and composite channel. In each case we quantify the relative increases in the shear stresses through time that are a direct result of the ice motion and parameter variation.

Proceedings of the Workshop on Environmental Aspects of River Ice, T.D. Prowse (Editor), National Hydrology Research Institute, Saskatoon, Saskatchewan, 1993, NHRI Symposium Series No. 12, p. 161-175.

INTRODUCTION

The movement of sediments can greatly alter the physical, chemical and biological environments of a river system. Bank and floodplain erosion affects the ecosystems of these areas, and can introduce toxic materials adsorbed on the sediments into the river. Bed scour and deposition affect the benthic communities and fish spawning habitat. Sediment transport creates an oxygen demand that can decrease the dissolved oxygen content of the water. High rates of sediment transport are generally associated with flood events. However, high suspended sediment loads have frequently been observed in rivers during ice breakup at a much lower discharge, but few systematic measurements are available.

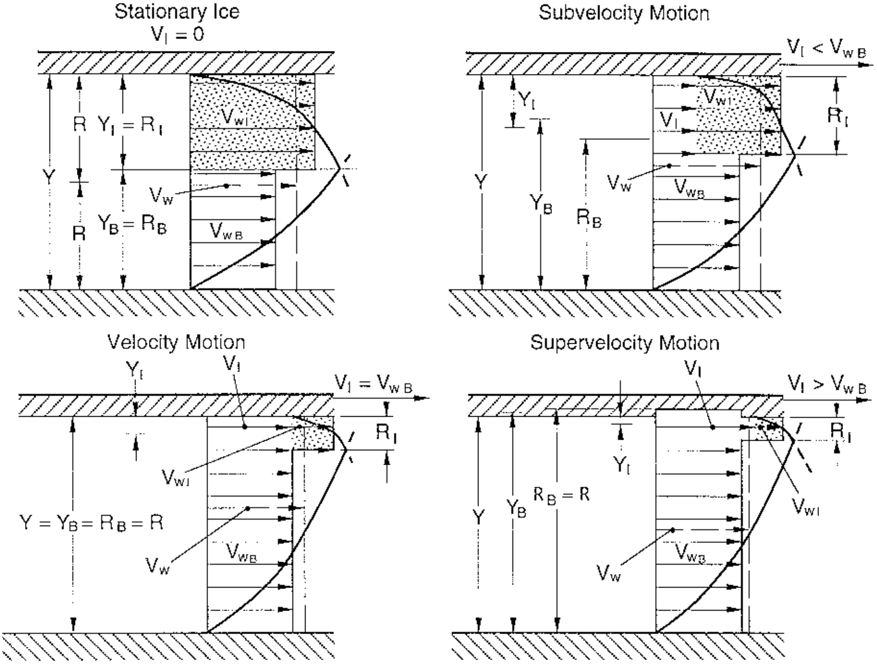
Prowse (1993) reported suspended sediment concentrations prior to and during breakup at one site on the Liard River in a single year. The concentration increased by an order of magnitude in the day following the initial ice motion, far exceeding that found during open water at the same flow. This preliminary work indicates that the transport during the breakup period could be a major contributor to the annual sediment budget of ice-covered rivers. Several mechanisms cause increased sediment supply and transport capacity during river breakup. Rapidly moving ice may directly scour the river banks and bed. Ice jams cause high water levels and locally high flow velocities. These high water levels expose additional bank areas to erosion, and cause flow diversion into erosion-susceptible floodplains. Dynamic river breakup exhibits rapidly rising river stage, velocity and discharge with high energy gradients immediately before and during ice motion. These flow surges can have a high sediment transport capacity.

Henderson (1966) cites the bed shear stress as the important measure of the power of the flow to dislodge and transport sediment. Yang (1984) summarized the work of several authors who assumed that bedload transport is directly related to the shear stress on the bed or to stream power, a related parameter. In the following analysis we assume that the sediment transport capacity of an ice-covered river is also directly related to shear stress. The quantity and size of the material mobilized on the bed is related to the bed shear stress, while the transport capacity of material suspended in the flow varies with the composite shear stress and the ice shear stress contributes to the downstream force on the ice. The three-velocity model of Ferrick and Weyrick (1993) is used to quantify the changes in shear stress caused by ice motion and parameter variations. A case study of breakup on the Connecticut River is generalized to examine the effects on the shear stresses of model input parameters that may be uncertain, including the relative roughnesses of the ice and bed, the ice velocity and acceleration, the flow velocity prior to ice motion, and the flow energy gradient.

MODEL DESCRIPTION

In this section we briefly describe the three-velocity flow resistance model of Ferrick and Weyrick (1993) for channels with a moving ice cover. We will use this model to quantify the changes in shear stress or sediment transport capacity caused by ice motion during breakup. The results of the model are internally consistent and closely follow physical intuition, but have not yet been verified by measurements. The river channel is idealized as wide and rectangular, and is treated as a composite, with the flow subdivided into portions that are resisted by either the ice or the bed. Composite parameters are obtained as a weighted average of the corresponding ice zone and bed zone parameters. The three average flow velocities of the ice zone, bed zone and composite channel are denoted V_{wI}, V_{wB}, V_w, respectively, and the ice velocity is V_I. The flow depths of the ice and bed zones, Y_I, Y_B, are the depths required to pass the flow resisted by the ice and the bed at the mean velocity of each zone. The total flow depth Y is measured to the bottom of the ice. The hydraulic radii of the ice and bed, R_I, R_B, define the zones of ice or bed resistance. R_I is the distance from the ice to the point of maximum velocity. The zonal depths and hydraulic radii are equal for stationary ice, but differ when the ice is in motion. The hydraulic radius of the composite channel is R. All of these model parameters are depicted in Figure 1 for both stationary and moving ice. The logarithmic velocity profiles in Figure 1 are generally accurate, but do not satisfy boundary conditions and have error near the intersection point and the walls (Chiu et al. 1993).

Figure 1. Depth, hydraulic radius and velocity parameters of the 3-velocity model for stationary ice cover, subvelocity, velocity and supervelocity ice motion. Light shading indicates the part of the velocity profile resisted by the ice. The intersection of the logarithmic profiles from the bed and the ice defines the point of maximum velocity in the model.



The three-velocity model contains four coupled equations in the unknowns R_I, R_B, V_{wI} and V_{wB}. Resistance equations for the ice and bed zones can be chosen from among Manning, Chezy, Darcy-Weisbach, integrated Prandtl-von Karman or other forms that relate flow velocity to hydraulic radius with a resistance coefficient. Resistance in each zone depends on the relative velocity between the corresponding boundary and the mean flow. Only the portion of the flow velocity greater than that of the ice is resisted in the ice zone. As a result, a V_I, V_{wB} compound profile must be resisted by the bed. The third equation replaces this compound profile with a single velocity V_{wB} extending over R_B using the requirement of equal momentum. The assumption of logarithmic velocity profiles from the ice and the bed with a common maximum at the intersection at R_I below the ice provides a fourth equation. With these parameters known, the composite velocity, the composite hydraulic radius and all shear stresses can be obtained.

Dimensionless parameters that represent ratios of like quantities are useful for comparing related flow conditions and model results. Ferrick and Weyrick (1993) define regimes of ice motion with a dimensionless velocity ratio

$$\tilde{V}_{rB} = \frac{V_I}{V_{wB}} \tag{1}$$

where the ~ indicates a dimensionless variable. In "subvelocity motion" $\tilde{V}_{rB} < 1$, indicating that the ice is moving more slowly than the mean water velocity in the bed zone. As the ice velocity increases relative to the water velocity we eventually attain a "velocity motion" transition with $\tilde{V}_{rB} = 1$. Finally "supervelocity motion" $\tilde{V}_{rB} > 1$ may occur with an ice speed greater than the mean flow velocity in the bed zone. All of the hydraulic parameters of the model respond continuously to changes in $\tilde{V}_{rB}$. A corresponding dimensionless velocity ratio for the ice zone is

$$\tilde{V}_{rI} = \frac{V_I}{V_{wI}} \tag{2}$$

During river breakup the changes with time in the bed load and suspended sediment transport capacities are directly related to the bed and composite shear stress changes, respectively. The shear stresses are expressed as

$$\tau = \rho g R S_f \tag{3}$$

where ρ is the water density, g is acceleration due to gravity, S_f is the flow energy gradient, assumed equal for each zone and the composite channel, and R is either the ice, bed or composite hydraulic radius corresponding to the related shear stress. Equal dimensionless shear stresses and hydraulic radii follow from the following definition:

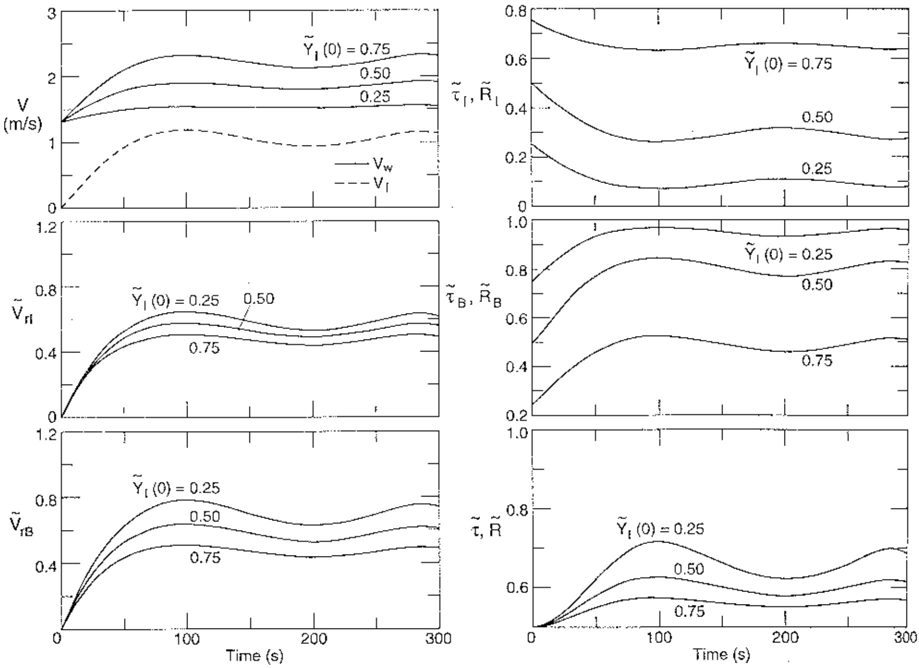
$$\tilde{\tau} = \frac{\tau}{\rho g Y S_f} = \frac{R}{Y} = \tilde{R} \tag{4}$$

Shear stress results that are presented in dimensionless form allow more general interpretation of relative magnitudes and percentage change in magnitude of the corresponding sediment transport capacity.

SENSITIVITY ANALYSIS OF THE SHEAR STRESSES

The ice velocity polynomial as a function of time, presented in Figure 2, is based on data measured at one location by Ferrick et al. (1992) during a dynamic breakup of the Connecticut River. River stage was also measured with time at the

Figure 2. Composite velocity, ice velocity, dimensionless velocity ratios (ice, bed), dimensionless shear stresses or hydraulic radii (ice, bed, composite) through time for a range of dimensionless initial ice zone depths.



same location, and ice thickness was known. We will examine the effect of ice motion together with that of systematically varied model input parameters on the sediment transport capacities of a river. These parameters are the relative roughness of the ice and bed, ice velocity and acceleration, flow velocity just before ice motion, and energy gradient during the motion. Manning resistance with constant bed and ice roughness through time is used in all simulations because of familiarity and to simplify interpretation of the results.

The relative ice-bed roughness conditions were varied over a range while keeping the same composite flow resistance prior to ice motion. The dimensionless ice zone depth prior to motion $\tilde{Y}_I(t=0) = Y_I(0)/Y(0)$ is the parameter that defines the relative boundary roughness, ranging from smooth ice with $\tilde{Y}_I = 0.25$, to rough ice with $\tilde{Y}_I = 0.75$. The initial composite flow velocity presented in Figure 2 is the same in all cases and the modeled velocity curves have the same basic shape. However, as $\tilde{Y}_I$ increases, the amplitude of the flow velocity response to the ice motion increases and the changes occur more rapidly. The dimensionless bed zone velocity ratio indicates subvelocity motion in all cases. The bed zone flow velocity is more strongly affected by changes in relative ice-bed roughness than the ice zone flow velocity. The dimensionless shear stress and hydraulic radius of the ice decrease following the start of motion, and the residual $\tilde{\tau}_I$ at later times increases with ice roughness. The response of the corresponding dimensionless bed parameters is the opposite. The amplitude of the change in suspended sediment transport capacity represented by the dimensionless composite shear stress and hydraulic radius increases as the relative ice roughness decreases, similar to the velocity ratios.

The sensitivity of the dimensionless shear stresses to ice velocity and acceleration was determined for the data given in Figure 3. These data were obtained by multiplying the ice velocity polynomial by 3/2, 1, 2/3, respectively, representing generally increased or decreased ice velocity and acceleration. Figure 4 depicts the velocity and shear stress results for both rough and smooth ice conditions with other parameters unchanged. The effect of these ice velocity

Figure 3. Ice velocity polynomial developed from data measured on the Connecticut River and related curves obtained from this polynomial with (3/2, 2/3) multipliers.

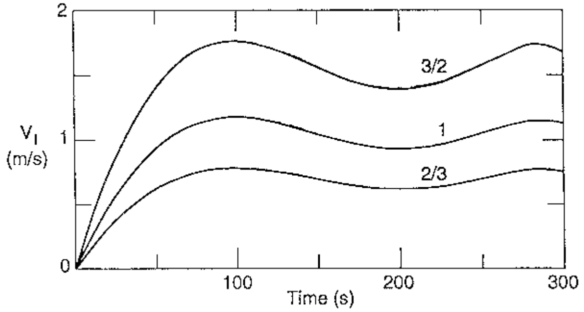
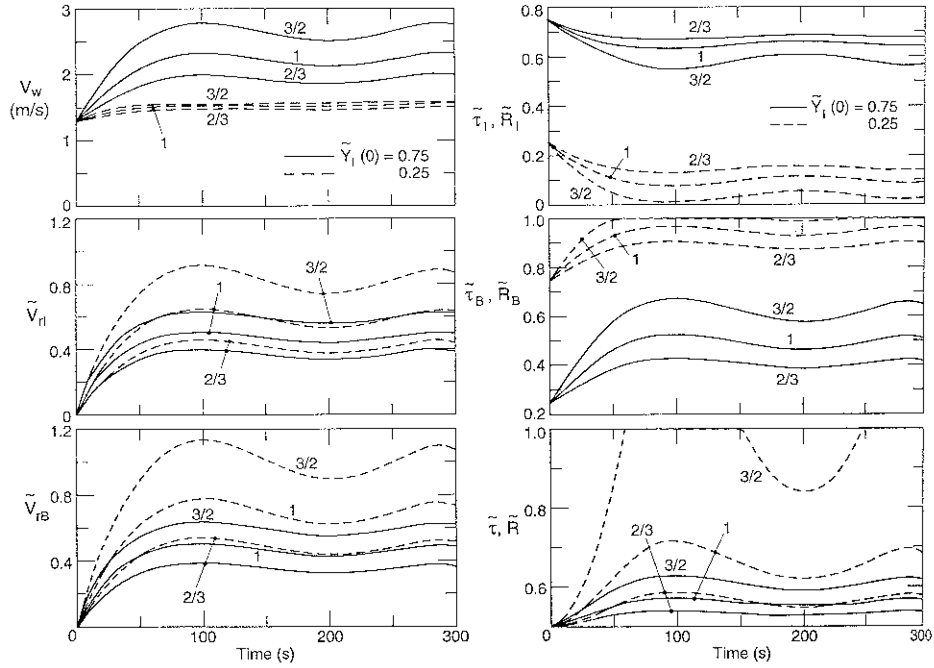


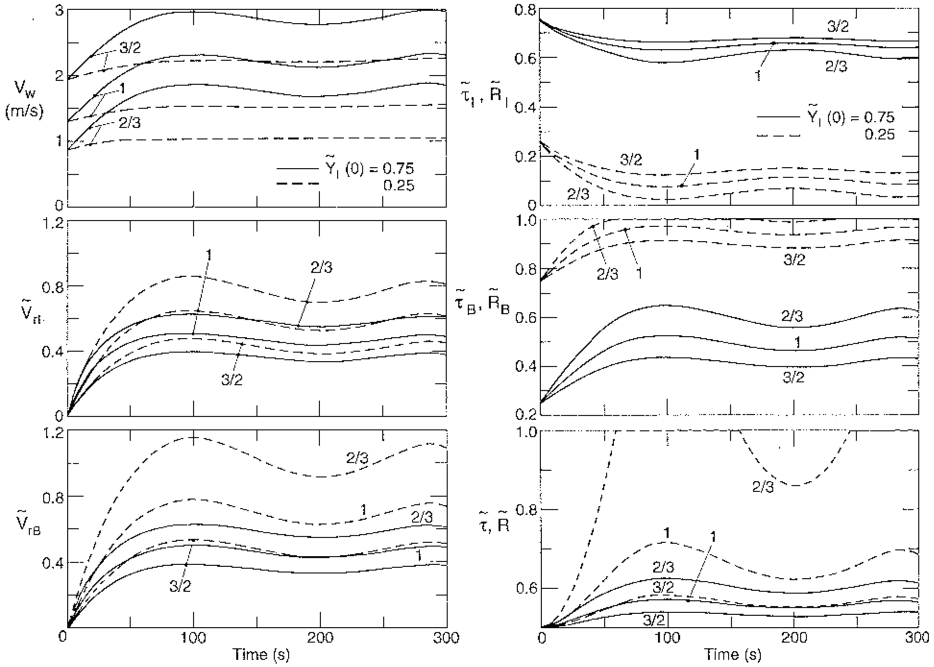
Figure 4. Composite velocity, dimensionless velocity ratios (ice, bed), dimensionless shear stresses or hydraulic radii (ice, bed, composite) through time for ice velocity multiples (2/3, 1, 3/2) and a pair of dimensionless initial ice zone depths.



differences on composite flow velocity is minor for smooth ice, but significant for rough ice. The dimensionless bed zone velocity ratio increases with the ice velocity and with decreasing ice roughness, and indicates supervelocity motion for part of the simulation in one case. The amplitude of the bed zone velocity response to changes in ice velocity is smaller than that of the ice zone velocity for smooth ice, and comparable for rough ice. The dimensionless shear stress on the ice is generally greater for rough ice as before, and also increases with decreasing ice speed. The response of the dimensionless bed shear is again the opposite. The rapid motion of rough ice provides the greatest percentage increase in the bed load transport capacity. The suspended sediment transport capacity follows the dimensionless composite shear stress and hydraulic radius, generally increasing with ice speed and with decreasing ice roughness.

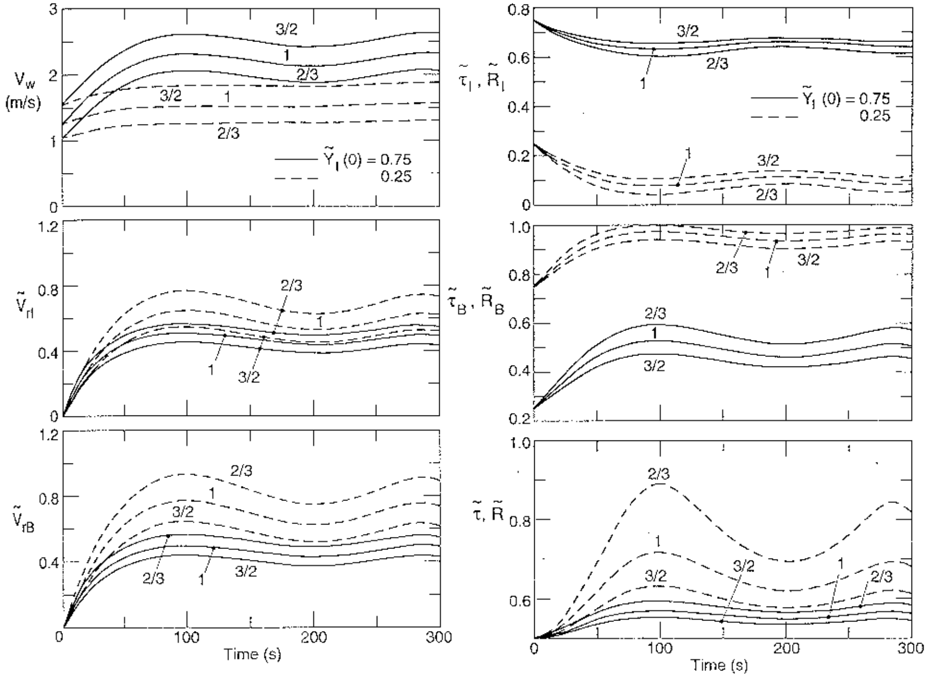
The initial flow velocity and discharge prior to ice motion may not be known precisely. We applied multipliers (2/3, 1, 3/2) to the initial flow velocity for both rough and smooth ice with other parameters unchanged to investigate the sensitivity of the sediment transport capacities to this parameter. The composite flow velocity responses to ice motion given in Figure 5 have an offset that is caused by the initial condition. With this exception, the changes in composite velocity through time are only weakly affected by the initial velocity. The dimensionless bed zone velocity ratio increases with decreasing initial velocity and ice roughness. A part of the simulation in the most extreme case developed supervelocity motion. The amplitude of the bed zone velocity response to changes in the initial composite velocity is smaller than that in the ice zone for smooth ice, and comparable for rough ice. The dimensionless shear stresses on the ice increase with the initial velocity, but ice roughness changes produce a much larger effect. The variability of the dimensionless bed shear stress with the initial composite velocity is much greater than that of the ice. The amplitude of the dimensionless composite shear stress change increases with decreasing initial velocity and ice roughness. Decreased initial flow velocity corresponds to generally increased bed load and suspended sediment transport capacities following ice motion.

Figure 5. Composite velocity, dimensionless velocity ratios (ice, bed), dimensionless shear stresses or hydraulic radii (ice, bed, composite) through time for initial flow velocity multiples (2/3, 1, 3/2) and a pair of dimensionless initial ice zone depths.



Because the energy gradient may not be known precisely, we applied multipliers (2/3,1,3/2) to the constant energy gradient for both rough and smooth ice with other parameters unchanged. For unchanged depths and roughnesses, the flow velocity was multiplied by the square root of the energy gradient multiplier. The results presented in Figure 6 are qualitatively similar to those for the initial flow velocity. Except for an initial offset the composite velocity responses to ice motion are only weakly affected by the energy gradient variations. The dimensionless bed zone velocity ratio increases with decreasing energy gradient, initial

Figure 6. Composite velocity, dimensionless velocity ratios (ice, bed), dimensionless shear stresses or hydraulic radii (ice, bed, composite) through time for energy gradient multiples (2/3, 1, 3/2) and a pair of dimensionless initial ice zone depths.



velocity and ice roughness, and indicates subvelocity motion in all cases. The amplitude of the bed zone velocity response to changes in the energy gradient is smaller than that of the ice zone for smooth ice, and comparable for rough ice. The dimensionless shear stresses on the ice increase with the energy gradient. For rough ice, the amplitude of the changes in dimensionless bed shear stress are much greater than those of the ice shear stress. The amplitude of the dimensionless composite shear stress changes increase with decreasing energy gradient and ice roughness. The changes in the bed load and suspended sediment transport capacities resulting from ice motion are generally increased for smaller energy gradients.

CONCLUSIONS

Ice velocity, thickness and roughness, plus standard stream gauging measurements provide input to our flow resistance model for channels with a moving ice cover. The flow resistance is coupled to the ice motion in the model, providing time variable mean flow velocities, peak flow velocity, discharge, and shear stresses on the ice, the bed and for the composite channel. These shear stresses determine the bed and suspended sediment transport capacities of the flow. We used this model to examine the sensitivity of the shear stresses to ice motion with systematic variations of measured input parameters, and presented the results in dimensionless form.

Changes in the motion of a river ice cover are generally accompanied by and contribute to surging flow. The sensitivity analysis performed using the model indicated that: 1) rough-bed-smooth-ice conditions produce a generally increased bed load transport capacity with ice motion relative to corresponding smooth-bed-rough-ice, 2) high velocity and acceleration of rougher ice cause the greatest percentage increase in the bed transport capacity along with a greater surge in the flow velocity and discharge, 3) the amplitude of the increase in suspended sediment transport capacity grows with increasing ice velocity and acceleration, and with decreasing relative ice roughness, 4) higher initial flow velocity and generally higher energy gradients lead to generally higher flow velocity and discharge, but smaller increases in the transport capacities due to ice motion.

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DISCUSSION

Edward Chacho:

How is the presented velocity and shear stress theory applied to sediment transport? If a standard shear stress transport equation is used, how do you account for movable bed when the high shear stress increases are encountered? (Shear stress transport equations are difficult to apply, even in open channels.)

Reply:

In this preliminary work, we assume that the bed load and suspended sediments transported are directly related to corresponding shear stresses. Modelling the sediment motion response to these stresses has been done by others and is not the intent of this paper. Prior to our work, a method to quantify the relationships between flow velocity, discharge, shear stress, and ice motion was not available, and we are attempting to fill this void. The model can accommodate a moveable bed with roughness changes through time if these values can be calculated provided as input. The ice roughness, a measured input to the model, generally changes with time, and bed roughness changes can be treated similarly.

Maurice Sydor:

Is the model developed for a single channel or networks? What are the special features of the cross-section being studied and is the "real-case" similar to those theoretical assumptions?

Reply:

The model calculations apply to a particular location on a river such as at a stream gauging station, where the requisite measured data are available. Multiple channel networks could be analyzed if data were obtained for each channel. In this preliminary study, the cross section for which calculations were made was taken for simplicity as rectangular. This idealization is close to the actual cross section. An arbitrarily shaped section could be considered at the expense of additional algebra.

Gee Tsang:

Physically I have difficulty in seeing how an ice floe can flow faster than the carrying water. Can you explain that?

Reply:

If an equation of motion written for an ice floe contains negligible resistance to ice motion, the resulting ice speed can exceed the mean velocity of the water. In these 3-velocity model simulations, the ice speed sometimes exceeded the mean water speed in the bed zone, but it never exceeded the mean water speed in the adjacent ice zone. The ice always resisted the motion of the adjacent water.