DISCHARGE MEASUREMENT DURING RIVER ICE BREAK-UP

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ABSTRACT

Both shore-based and aerial survey techniques for measuring surface ice velocities are reviewed. Small format photography is suggested as the most practical method of aerial monitoring. Results of surveys using a recently developed 35mm system are described. Methods and potential problems of estimating river discharge from surface ice velocities are detailed.

INTRODUCTION

The standard practice for discharge monitoring in Canada is based on establishing a reliable stage-discharge relationship. The procedures are relatively straight forward under open water conditions, but complications arise when the flow is affected by ice. Because of the additional flow resistance created by the ice, water levels are often elevated over those for equivalent discharge under open water conditions. If a stable ice cover develops, discharge measurements can be periodically made through the ice and then, by various methods, water level records can be converted to discharge (see for example, Rosenberg and Pentland 1966; Santeford and Alger 1986).

More severe obstacles to standard measurement procedures occur during freeze-up and break-up when ice conditions are more variable and no consistent relationship can be established between stage and discharge. The ice cover is usually too unstable or weak to traverse in order to perform on-ice discharge measurements, and even if a bridge or cableway exists, there is often too much ice moving in the flow to permit the lowering of current meters. These conditions may persist for days or even weeks. On the Mackenzie River near Fort Simpson NWT, for example, the average time between the appearance of the first permanent ice and complete freeze-over is one month, and the usual time between the first deterioration of ice and clear-of-ice conditions is two weeks (Allen 1977). Thus, conventional measurement techniques would be out of the question at the Mackenzie site for approximately six weeks each year. To make matters worse, in some regions of Canada multiple freeze-up/break-up events can occur in a single winter, prolonging the period during which measurements are not logistically feasible.

In drainage basins where break-up progresses downstream, a standard practice is to estimate mean daily discharge by extrapolating flow data from upstream ice-free sites to downstream ice-affected stations. Allowances are made for flow travel time and hydrometeorological events such as snowmelt and rainfall. A major drawback of this approach is the difficulty in adequately accounting for rapid fluctuations in flow and water levels which occur on time frames of much less than one day. Gerard (1979) recounts one event in which water levels increased over 17 m in less than one hour. In most cases, such rapid changes can be attributed to surges resulting from the formation and subsequent failure of ice jams (Henderson and Gerard 1981, Beltaos and Krishnappan 1982). Ferrick (1985) notes that ice jams can produce long-period waves in rivers, and remarked upon the significance of the waves for river ice break-up. Several recent studies, such as that by Andres and Doyle (1984) for the Athabasca River and Prowse (1984) for the Liard River, point to the surge-stall behaviour of break-up on northward-flowing rivers, a characteristic attributable to repetitive ice jam formation and release.

Monitoring of instantaneous discharge during the freeze-up and break-up periods is a key to providing answers to a number of winter flow problems, particularly those related to ice jams. For example, Andres and Doyle (1984) outlined a solution to the equilibrium jam formulation which requires instantaneous discharge measurements at the time of jam formation and a short time later after water levels have receded. Given the unsteady flow conditions which prevail during such periods, estimates of mean daily flow are of limited use in such analyses. Alternative methods of acquiring frequent estimates of instantaneous discharge are clearly needed. This paper reviews some of the remote-sensing techniques which have been developed for this purpose and details some of the problems which may be encountered.

APPROACHES

Most of the remote sensing techniques involve three steps: a) measurement of surface ice velocities, b) correction of these to mean velocity values and c) a discharge calculation using information about ice distribution and flow depth.

Shore Survey - Trigonometry

One of the simplest approaches (Fig. 1) involves tracking individual ice floes from shore using a theodolite (see, for example, Michel 1971). During installation the distance above the water surface to the theodolite is surveyed. Velocities are to be measured at a number of locations across the cross-section and therefore the distance away from shore must be known. By simple trigonometry, this can be calculated from:

[1] $b_i = d_E/\tan \alpha - b_S$

where b_i is the shore to ice floe distance (m), d_E the elevation of the survey point above the water surface (m), α the vertical angle to the ice floe measured perpendicular to the flow and b_S is the horizontal distance from the water's edge to the survey point (m).

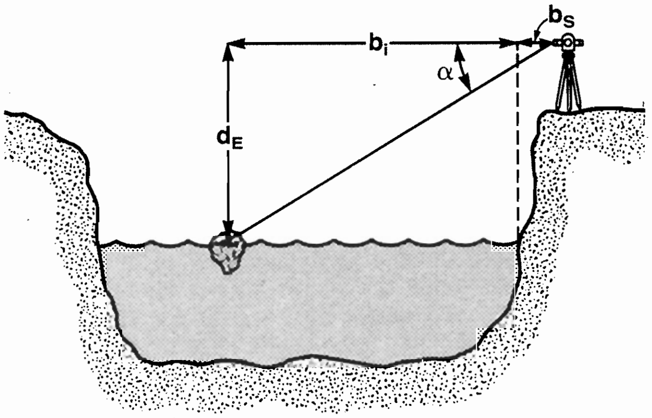

Figure 1. Shore survey of ice floe location.

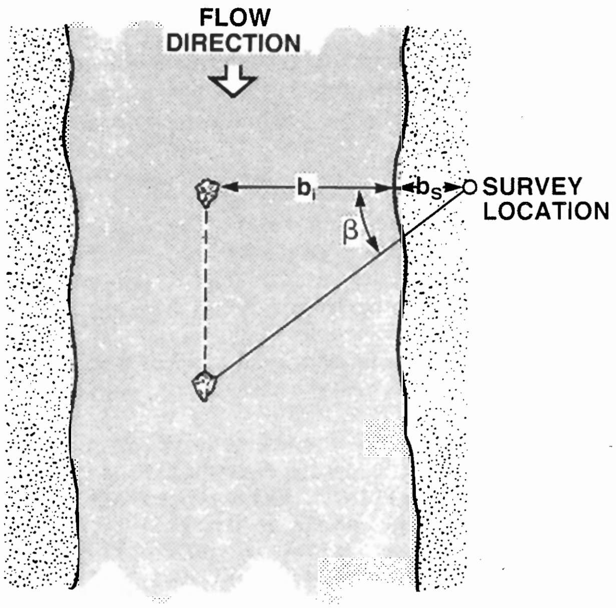

Figure 2. Shore survey of surface ice velocity.

The velocity of surface ice floes can be determined (Fig. 2) by tracking a floe over a short distance from a point perpendicular to the shore to a downstream location. The surface ice velocity is then given by:

$$[2] \qquad v_i = (b_i + b_S) \tan\beta / t$$

where v_i is the surface ice velocity (m/s), β the horizontal angle of movement, and t is the elapsed time (s). Ice concentration can also be determined with this technique by calculating the percentage of time over a fixed interval that ice occupies one of the observation points.

<u>Shore Survey- Photographic Grid</u>

A more elaborate shore-based technique for obtaining surface ice velocities is used by the Hungarian Water Conservation Bureau (1980). Camera platforms are located on the banks at a height of approximately 0.1 times the width of the cross-section to ensure adequate coverage and good photographic definition for later analysis. For wide rivers, this may not be practical and lower camera elevations become necessary. On the opposite bank, signboard markers are erected perpendicular to the optical axis of the camera and at a set distance apart. Another two temporary markers are established on the near shore separated by a similar distance. A reference photograph is then taken which encompasses all four targets, thus forming a rectangle. Using this reference photograph, a grid can be constructed such as that shown in Fig. 3. Following Canadian standards, the lines parallel to the flow direction should be spaced at intervals equal to approximately 5% of the river width, although this may not be possible where camera elevations above the water surface are restricted. Lines perpendicular to the flow should be approximately 10 m apart.

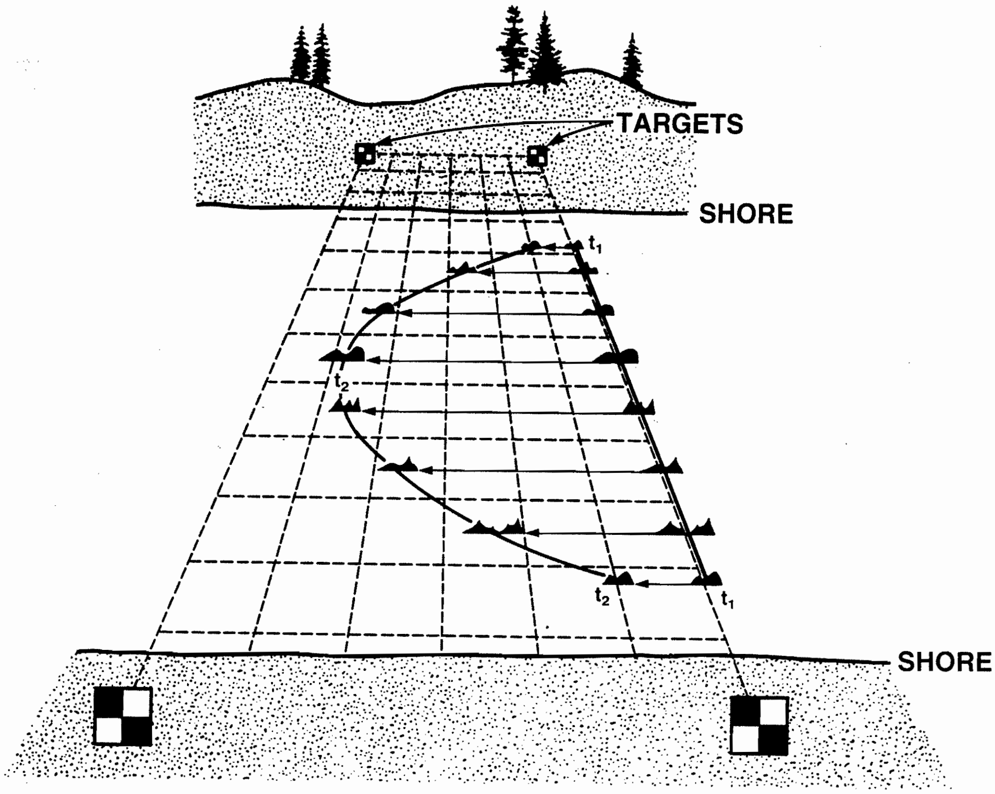

Figure 3. Photographic grid for determining ice velocity.

During freeze-up and break-up, time lapse photography is taken of ice moving through the section. The elapsed time between photographs must be less than the time required for the ice to pass between the pairs of targets. These photographs are then used with the reference grid photograph to determine both surface ice velocities and ice concentration.

Aerial Surveys - Floats

Airborne camera systems have also been used in the determination of discharge. In the Soviet Union, for example, a surface velocity technique which employs air-dropped floats and aerial photography has been under development since 1965 (Shumkov 1973). Some experimental work has also been done with hydrobombs which release oil through the water column and allow the determination of an integrated velocity (Kuprianov 1978).

Aerial Survey - False-parallax

For some time, it has been realized that aerial photographs can also yield information about the motion of objects through the creation of a false- or pseudo- parallax. Stereoscopic interpretation of photographs relies on parallax which is "the apparent displacement of the position of a body, with respect to a reference point or system, caused by a shift in the point of observation" (Thompson 1966, p. 1147). Displacement of the object rather than the point of observation can also produce a parallax (Fig. 4). Successive photographs of a moving object can thus be used to determine the object velocity. Where the camera movement and object motion are in opposite directions, the image of the object viewed stereoscopically will appear elevated above the adjacent stationary landscape. Higher elevations correspond with increased velocities of the object. Conversely, if the camera and object motion are in the same direction, the image will appear depressed relative to its surroundings.

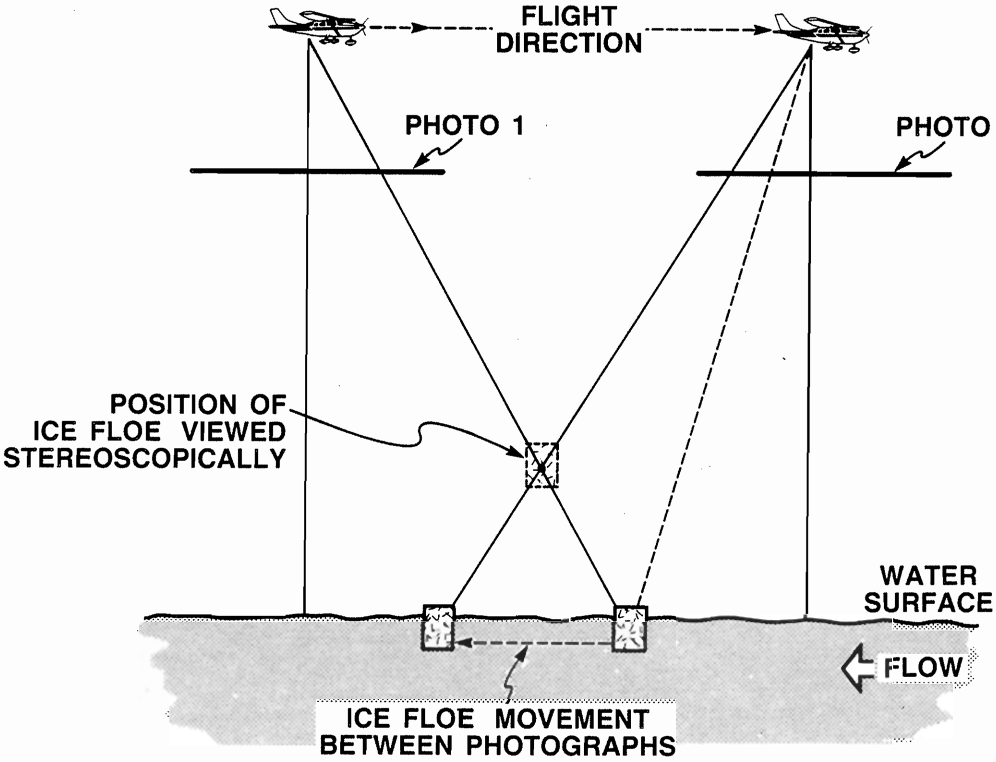

Figure 4. False-parallax produced by moving ice floe.

Cameron (1952,1962) first used the pseudo-parallax approach in the analysis of water currents. It has subsequently been used by, for example, Forrester (1960) in rivers, Duhaut (1972) and Keller (1975) in tidal zones, and by Moffitt (1968) for mapping the wake from ships. As noted by Cameron (1962, p.158) two conditions are necessary for the application of this technique:

"1. The water surface must be marked, either naturally or artificially, so that the same area or spot can be recognized stereoscopically.

2. Some stationary object or surface must be present in the stereo overlap to act as a zero reference when measuring the anomalous parallax by movement of objects or surfaces between the time of the two photos."

Ice floes within a river during freeze-up or break-up create ideal targets for this technique. Beginning in the early 1970's, the National Hydrology Research Institute (N.H.R.I.) of Environment Canada conducted annual aerial photographic surveys of river ice break-up on the Liard and Mackenzie Rivers. Some of this imagery was analyzed photogrammetrically to assess the operational potential of the pseudo-parallax technique (MacKay et al. 1974; Sherstone 1980).

Aerial Survey - Digital image analysis

One of the apparent drawbacks with the false-parallax approach was the lengthy time required to compute and map the surface velocities. A more automated process of image analysis was subsequently developed by N.H.R.I. (Langham et al. 1981). The photographs were optically digitized and statistical procedures were used to compare image shifts in the ice-water mixture and to determine surface ice velocities.

Aerial Survey - Small format photography

Large format (9") aerial camera systems, such as those used in the above studies, provide high resolution imagery, but the cost of acquiring the information is often prohibitive. Not only are the camera systems expensive to own or lease, but the associated operating costs such as air charter, film and processing are also high. As costs have escalated and small format (35 mm) cameras have become more sophisticated, increasing consideration has been given to conducting aerial surveys with the small format systems (see for example, Clegg and Scherz 1975, Sherstone 1978).

In response to the need for inexpensive aerial monitoring of river ice conditions, NHRI developed a 35 mm aerial camera system. Because of the high costs and lengthy time required to apply the false-parallax or image-shift analysis to 35 mm images, the system was designed for obtaining timed sequential aerial photography from which the displacement of ice floe images and therefore surface ice velocities could be directly measured. The system includes a 250-frame Canon F1 camera housed within an "Enviro Pod" designed by DIDEC Research and Development Ltd. (Fig. 5). The pod can be fastened beneath a Cessna 185 fixed-wing aircraft (Fig. 6) or Bell Jet Ranger helicopter. Christie and Maxin (1985) describe the control system which includes a frame timer, frame counter, and audio-visual indicators for monitoring camera operation (Fig. 7).

EVALUATION

The remainder of this paper describes the procedure and problems associated with using 35 mm aerial photography for determining discharge during break-up. Many of the problems identified are also common to the other techniques reviewed above.

Test Site

During the spring of 1985 the camera-pod system was used to record the advance of river ice break-up on the Liard and Mackenzie Rivers near Fort Simpson, NWT. A detailed description of the 1985 break-up is presented in Prowse (1986a) and an analysis of annual ice jam formation at the confluence of these two rivers has been completed by Prowse (1986b) for the years 1978-1984. During the break-up period, Water Survey of Canada makes estimates of mean daily flow, but there are no direct measurements or estimates of instantaneous discharge. Since there is no way to independently verify the accuracy of the results from the photographic surveys, discussion is limited to problems encountered in collecting and processing the ice velocity data.

When the weather and ice conditions were suitable the camera system was used to record ice velocities at two Water Survey of Canada gauging stations. Station No. 10ED002

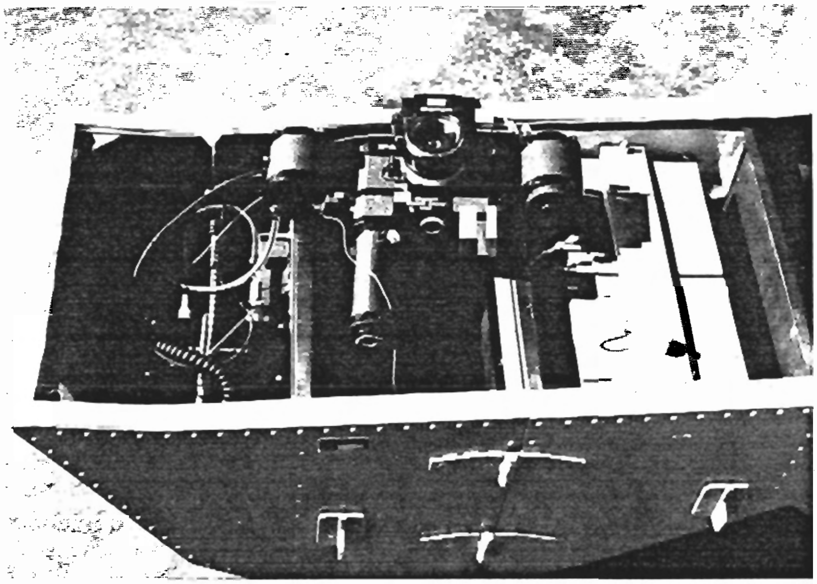

Figure 5. Canon F-1 35 mm camera within pod.

Figure 6. Camera pod mounted on Cessna 185.

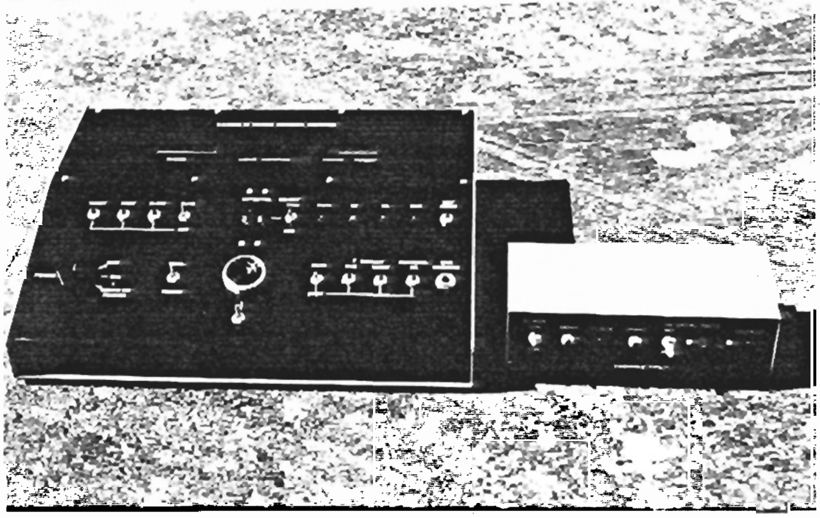

Figure 7. Camera timing and control system.

is located at the Liard River ferry crossing, 17 km upstream of the river mouth. The second station (No. 10GC001) monitors water levels on the Mackenzie River at Fort Simpson island, approximatley 2 km below the Liard River confluence. During the fall of 1984 and 1985, NHRI staff conducted bed and bank surveys at the two hydrometric stations and established the following relationships:

[3] $b = 6.11\ d_G + 693.92$

for station No. 10ED002 and

[4] $b = 13.11\ d_G + 1066.42$

for station No. 10GC001, where b is the channel width (m) and d_G is the Water Survey of Canada gauge height (m). (To convert gauge heights to Geodetic Survey of Canada datum add 110.71 or 115.90 m for the Liard and Mackenzie stations respectively.) Changes in water levels during the 1985 break-up are shown in Prowse 1986a. During much of the period, water levels and flow widths were significantly greater than those which normally occur during the open water period.

Data collection

The flying height to encompass the desired channel width is calculated from:

[5] $h = b\ f/l$

where h is the flying height above the ground (m), f is the focal length of the camera (mm) and l is the effective side length of the film frame (mm). This relationship can also be used to determine the scale of the photograph if elevation can be accurately measured, such as by radioaltimeter. The small aircraft used in this application, however, employ aneroid or pressure altimeters which provide only a coarse estimate of elevation and are insufficiently accurate for the determination of a photographic scale. It was therefore necessary to use ground references for scale calculations. At the Fort Simpson station, map surveys of town streets were used to scale the photography. The more remote Liard station required that 1 m^2 targets be located 100 m apart along a 0.5 km section of the left bank.

Wide-angle lenses (24 and 28 mm) were used because of the large channel widths. Longer focal lengths would have required flying heights at which photography was often impaired by low level clouds. Image tests performed on the wide-angle photography indicated minimal distortion. Errors in the discharge estimates arising from lens distortion and from standard commercial processing were considered insignificant.

One of the largest sources of error in aerial photography is due to the movement of the plane and camera during photography. Vertical aerial photographs are usually taken at angles of less than 3^o from the vertical and therefore have a uniform horizontal scale. However, with light aircraft operating in the often extreme turbulence at low elevations, it was impossible to ensure vertical photography. Many photographs were oblique ($>3^o$ to vertical) and therefore characterized by a variable scale. Since it was impractical to rectify the photography, photographs to be used in the velocity analysis were selected according to their apparent level of verticality. In most tests, the flying height was low enough that the river width almost completely filled the photographic image. Any photographs which did not contain a complete image of the river cross-section along the centre line of the photograph indicated either strong tilt or tip of the aircraft and were discarded. A detailed review of the various errors associated with oblique photography is contained in Shumkov (1973).

Velocity and discharge calculations

The surface velocity of an ice floe can be determined from:

[6] $v_i = P_S\ L/t$

where P_S is the photographic scale factor, L the downstream travel distance between photographs (mm) and t is the interval between photographs (s).

The surface ice velocity, however, does not necessarily reflect the surface water velocity. Movement of the ice floe is dependent on the forces imparted by the water current and wind flow which are defined as:

$$[7] \qquad F_A = p_A\, C_A\, A_A\, (U \cos \phi - v_i)^2/2$$

$$[8] \qquad F_W = p_W\, C_W\, A_W\, (v - v_i)^2/2$$

where the subscripts A and W refer to air and water respectively, F is the drag (N), p is density (kg/m^3), A is exposed area (m^2), C is a drag coefficient, U and v are the velocities (m/s) of the wind and water respectively, and ϕ is the angle (degrees) of the wind relative to the current direction.

Under normal ranges of wind and river water velocities the effect of water drag is much larger than that of wind drag. This is primarily due to the large density differences between air and water and because a much larger surface area is exposed to water than air due to the buoyancy of ice. To illustrate the relative importance of wind in the movement of ice floes, consider a break-up in which an ice floe has reached equilibrium with the wind and water forces. Assume a break-up water velocity of 1.0 m/s and a windspeed of 3 m/s which is the mean value for the study site during the break-up period (Fort Simpson A, AES 1980). Also assume similar drag coefficients and a ratio of $A_A{:}A_W$ equal to the volume ratio produced by natural buoyancy of ice (assumed density of 910 kg/m^3). In this example, the assumed windspeed from the climate station is representative of conditions 10 m above the surface. Given a logarithmic wind profile, much lower near-surface windspeeds would strike the small height of ice protruding above the water surface. A value of U = 3.0 m/s can therefore be considered representative of a high wind situation.

Substituting the above values and equating [7] and [8], results in $v/v_i = 0.97$. Thus, the force imparted by the wind accounts for only 3% of the ice floe velocity. For the windspeed to account for just 10% of v_i, the water velocity would have to be less than 0.3 m/s. In cases of large ice pans, rather than small diameter floes, calculation of forces would depend more on skin roughness rather than the form roughness considered above. A similar situation could develop where small diameter floes are tightly packed together such as in ice runs. The relative importance of the forces would, however, remain approximately the same.

Given the relatively small importance of wind drag, surface ice velocity is considered for practical purposes to be equal to surface water velocity. However, because of the variation of velocity with depth, the surface water velocity must be converted according to:

$$[9] \qquad \bar{v} = K\,(v_i)$$

where $\bar{v}$ is the mean velocity for a water column (m/s) and K is a coefficient which depends on the vertical velocity distribution. Table 1 lists K values for varying channel characteristics and flow depths.

Following the mid-section method for determining discharge the total river flow can then be determined from:

$$[10] \qquad Q = \sum_{i=1}^{n} \bar{v}_i\, d_i\, (b_{i+1} - b_{i-1})/2 + Q_L + Q_R$$

where Q is discharge (m^3/s), d_i the average flow depth (m) associated with the i-th ice floe), and b_i the cross section distance (m) from the initial point to the i-th floe. The initial point is determined as the mid-point between the shore and the first ice floe. Q_L and Q_R are the estimated discharges (m^3/s) for flow within the sections between the left bank and the first ice particle and between the right bank and the n-th ice particle, respectively.

Table 1. Value of K for different channel characteristics (after Kuprianov 1978).

Channel Characteristics	Mean Depth (m) <1	1-5	>5
a) Channels straight, clear, earthen, gravel.	0.80	0.84	0.86
b) Channels meandering, overgrown by grass, stony. Floodplains - well developed with vegetation.	0.75	0.80	0.83
c) Channels and floodplains -considerably overgrown with deep scours. Channels -meandering with large cobblestones.	0.65	0.74	0.80
d) Floodplains- completely covered by forest of the taiga type	0.57	0.69	0.75

Problems due to ice conditions

During river ice break-up the concentration of floating ice usually decreases from an initial situation of complete congestion to almost complete open water with a few scattered floes. To illustrate the range of problems associated with determining discharge during break-up, three types of ice conditions are reviewed. The first represents a period of heavy ice congestion (Fig. 8a) immediately following the initial fracturing and downstream movement of the Liard River ice cover. Once the initial ice run had passed the floe concentration rapidly decreased, fed only by ice from upstream tributaries and ice calving from shear walls along the shores. Fig. 8b shows conditions four days after the initial break-up. The ice concentration is <5% but the floes are relatively evenly distributed across the channel. The final situation involves an uneven distribution of ice in which scattered ice floes, originating from the Liard River, cover the left hand side of the Mackenzie River channel while most of the right side, fed by the upper Mackenzie River, is ice free (Fig. 8c). The floe locations used for analysis and the resultant velocity profiles are shown in Fig. 9. Preliminary results of the discharge calculations, using [6], [12] and [13], are listed in Table 2. As a first approximation, K was set to 0.8.

Table 2. Discharge analysis, Liard and Mackenzie Rivers, 1985. Symbols: n is ice floe sample size, V is mean floe velocity, σ_v is standard deviation of velocity measurements, other variables are defined in the text.

	d_G (m)	n	t (s)	V (m/s)	σ_v (m/s)	Q (m^3/s)
Liard River May 8 19:33 MST	9.92	19	240	1.21	0.09	8622
Liard River May 12 14:29 MST	9.46	20	300	0.98	0.10	5383
Mackenzie River May 12 14:53 MST	13.64	13	660	0.83	0.08	10531

Four basic problems associated with the break-up process were identified: a) selection of a representative K value, b) volume displacement by ice, c) width variations, and d) uneven floe distributions.

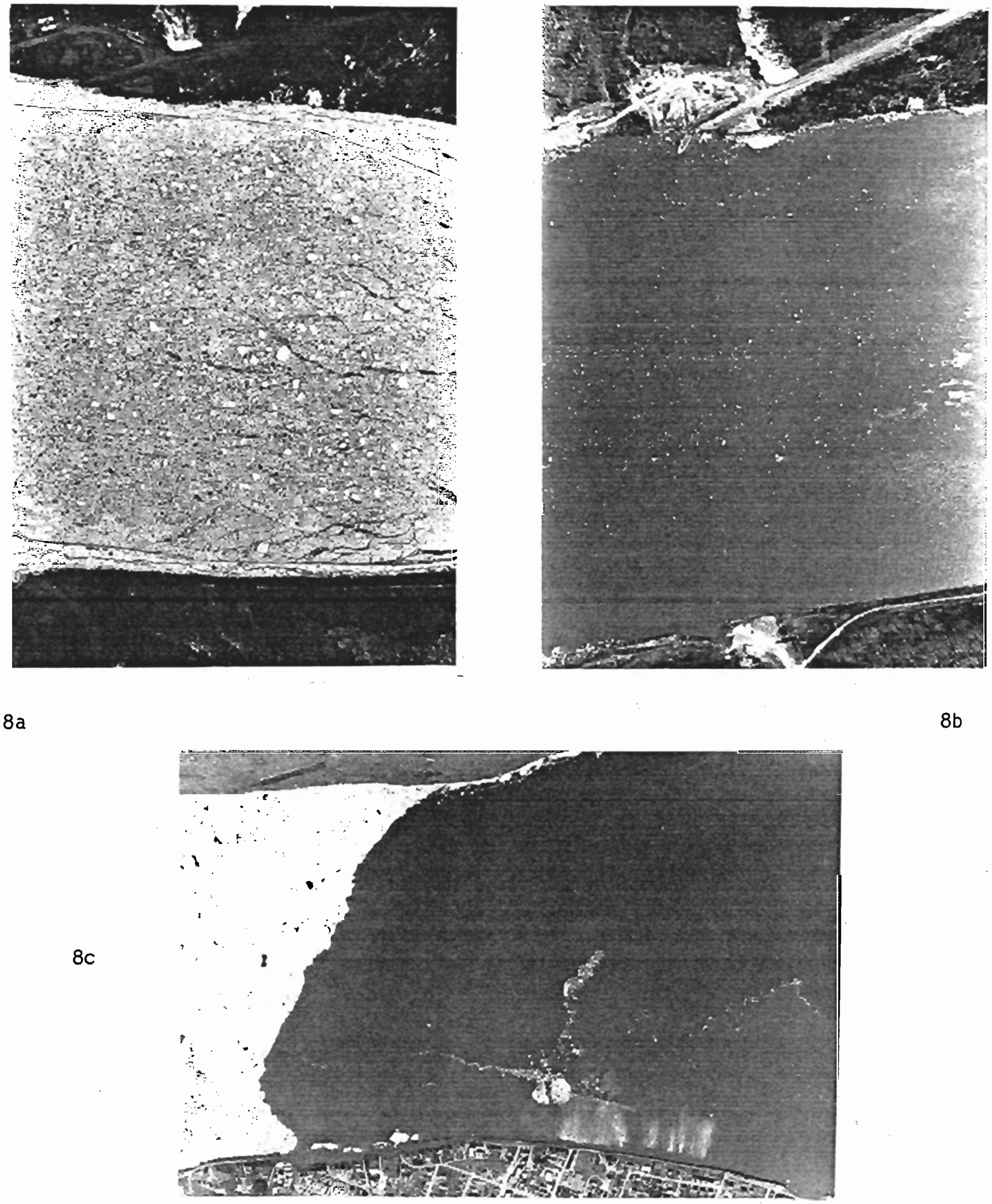

Figure 8. Ice conditions during break-up.
8a) Liard River May 8, 19:35 MST - immediately following initiation of break-up
8b) Liard River May12, 14:32 MST - <5% surface ice concentration, evenly distributed
8c) Mackenzie River May 12, 15:00 MST - uneven ice distribution.
Flow direction in all pictures is from right to left.

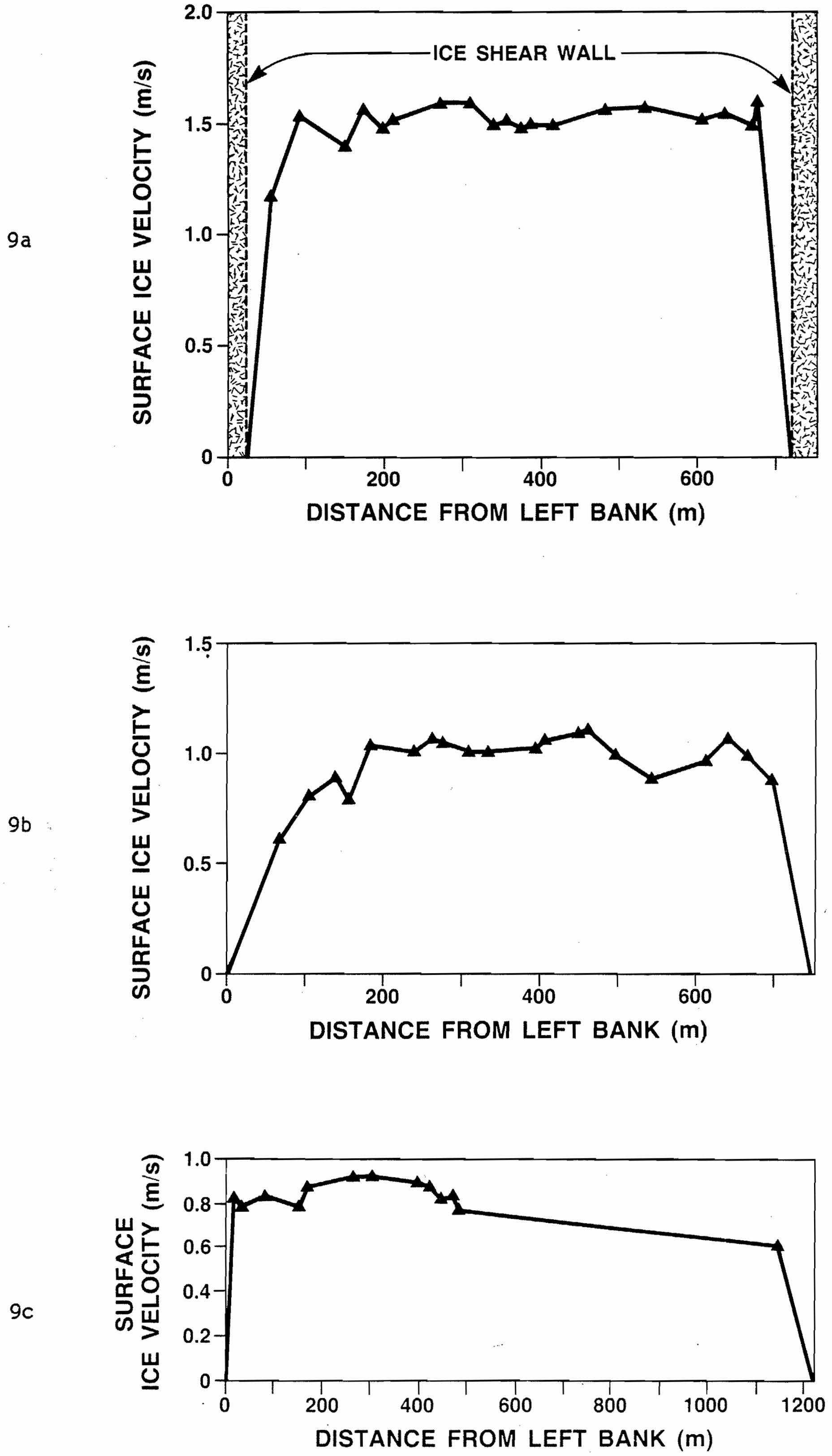

Figure 9. Surface velocity profiles related to ice conditions show in Fig. 8.
9a) Liard River May 8; 9b) Liard River May 12; 9c) Mackenzie River May 12.

a) K value

Selecting a K value, especially for conditions of high ice concentrations, is difficult because of the additional flow resistance created by the ice cover. The values of K in Table 1 are applicable to open water situations in which flow resistance is largely determined by the channel bed. The addition of a surface ice cover, even though it is moving, creates additional resistance to water flow and thereby reduces the mean velocity.

When ice concentrations reach a congestion point, in either the vertical or horizontal, additional resistance to ice movement results from contact with the shores and channel bed. In the vertical case, the probability of ice grounding on to the bed increases as the thickness of the ice accumulation approaches the flow depth. This can eventually lead to the formation of a "grounded" ice jam. Similarly, when the surface ice concentration is high enough that ice floes become interconnected, the banks create additional resistance to movement of the ice accumulation. At some point, the floes will bridge the entire section and initiate a jam. As conditions approach a state of jamming, the ice floe velocities become increasingly less representative of water velocities and assignment of a K value becomes more difficult.

b) volume displacement

When the channel is filled with a considerable volume of moving ice, the observed stage cannot be directly used to determine the flow depth for use in "liquid" water discharge calculations. Assuming a near rectangular channel, an adjustment can be made to account for the volume of ice according to:

[11] $$d = d_o - s_i t_i C_i$$

where d is the mean depth of flow without the ice cover (m), d_o is the observed flow depth (m), s_i is the specific gravity of ice, t_i is the mean ice thickness (m), and C_i is the surface concentration of ice (0.0-1.0). This correction can be considerable when the ice accumulation is thicker than a single floe, such as during periods of large ice runs resulting from the failure of ice jams. Measurement of t_i during such periods, however, is extremely difficult and must usually be estimated from the heights of shear walls formed by the ice run.

c) width variations

The formation of shear walls also poses a problem in the location of ice floes with respect to the bed profile. In the absence of shear walls, the method is straightforward. A ground scale is derived from the photographs, and channel width is then calculated and compared to that predicted by [3] or [4] using the stage records. Floe locations with respect to the shore locations are then determined and flow depths are obtained from the bed profiles. However, if large masses of stranded ice line the banks, a bank-water interface cannot be identified and ice floes cannot be accurately located with respect to the bed profile. Working from the left bank, approximate locations can be determined using (Fig. 10):

[12] $$b_L = S_L (b - b_o)/(S_L + S_R)$$

where b_L is the distance (m) between the shore-water and shear wall-water interfaces on the left bank, b is the channel width (m) determined from [3] or [4] which includes portions of the shear walls immersed in water, S_L and S_R are the respective widths of the left and right shear walls (m), and b_o is the channel width (m) between the shear walls as determined from the photography and ground scale. An assumption is made that the widths of the shear walls in water are proportional to the respective widths of S_L and S_R. Knowing b_L then permits positioning of b_o and the ice floes with respect to the bed profile.

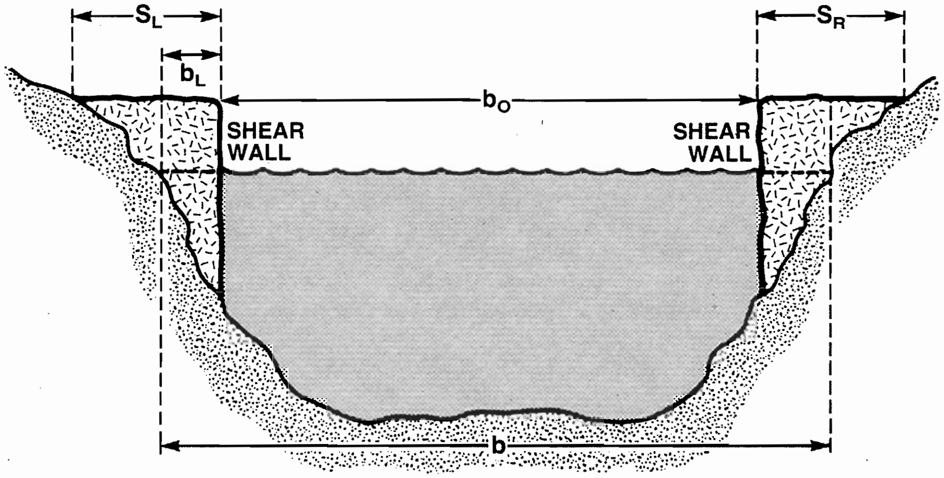

Figure 10. Ice shear walls and channel geometry.

d) floe distribution

During the latter stages of ice clearance, only scattered ice floes remain and these are frequently concentrated within specific sections of the channel. This leaves large open water sections in which velocities must be extrapolated over large river widths, well beyond the limits deemed acceptable under open water conditions. For example, flow estimates for approximately 60% of the Mackenzie River channel width in Fig. 9c are based on the velocity of two ice floes. The only remedy in such situations is to use artificial floats until open water discharge measurements can be resumed.

SUMMARY

Where observations must be conducted from river banks, trigonometric surveys are the simplest and most direct method for obtaining surface ice velocities. When measurements of instantaneous discharge are required, however, the lengthy time required to complete a full set of cross-sectional surveys reduces the usefulness of the technique, especially on wide rivers. Instantaneous discharge can best be measured using photographic shore surveys based on a reference grid-image. This approach is also one which could be easily integrated with existing hydrometric station networks to permit automatic recording during the freeze-up and break-up periods. One drawback of the approach for use in low relief areas is the large camera elevations required for wide rivers.

Large format cameras and the associated false-parallax and image-digitizing techniques of photogrammetric analysis form the most sophisticated and accurate methods for determining surface ice velocities. However, given the range of problems and accuracy limitations associated with converting surface ice velocities to discharge values, the accuracy levels of such sophisticated techniques are essentially surplus to the needs of this type of environmental mapping. Images produced by 35 mm photography are fully adequate and offer the opportunity of obtaining more surveys for the same cost. This is especially important considering the need for frequent surveys to define the rapid fluctuations in discharge which occur during the break-up period.

Problems associated with the scale and positioning of 35 mm images can be reduced if the channel bed profile can be accurately located relative to identifiable ground control stations. A minimum of three stations, preferably spaced widely apart, would permit rectification of the image and proper positioning of the shear walls and ice floes. For many locations where this approach could be used, surface maps of sufficient detail already exist. The only step that may be required is to accurately locate the bed profile with respect to the selected ground control stations.

The major problem in deriving discharge from surface ice velocities occurs during the periods of high ice congestion when the friction created by the banks and bed become important. Although K can be reasonably approximated during periods of low to medium ice concentrations, surface ice velocities cannot be considered representative of surface water velocities above these concentrations.

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