HUDSON RIVER ICE MANAGEMENT

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ABSTRACT

An ice management strategy is being developed for a reach of the Hudson River that experienced ice jam flooding during the 1983-84 winter. Preliminary field studies have focused on developing a technique to induce the breakup of an ice cover or ice jam by releasing water from an upstream dam. During these studies, a series of abrupt releases generated long-period river waves of different magnitudes, durations and spacings that caused changes in river level, flow velocity, and integrity of the ice cover. By monitoring the river elevation and ice cover at several locations, we have found that each of these wave parameters affected the response of the ice cover. The steepness of the wave front depends upon the initial river stage and the amplitude of the release, and is an important parameter affecting the stability of the ice cover. The sequence of events leading to breakup of the relatively thin ice cover on the Hudson was identical to that reported for other rivers having different physical characteristics and much thicker ice. These studies have revealed that pulsed releases of a practical magnitude were effective in removing the ice cover from the reach and provided basic data for analysis of river ice cover breakup.

Introduction

Early in the winter of 1983-84 a massive ice jam formed on the Hudson River near the towns of Hadley, Lake Luzerne, and Corinth, New York. The jam remained in place for over two months and throughout this time presented an immediate flooding threat to hundreds of residences. In the summer of 1984, we began a study of the factors leading to this ice jamming problem and of methods for minimizing future jams in this location.

Our study reach of the Hudson River (Fig. 1) extends from Rockwell Falls at Hadley-Lake Luzerne to Curtis Dam at Corinth, a distance along the river of 5.6 mi. (9.0 km). Backwater from the dam causes widening and deepening of the study reach relative to the river upstream, reducing the energy gradient of the flow. The energy gradient reduction, together with some obstruction to ice movement such as an existing ice cover, causes moving ice to jam. The Sacandaga River, a major tributary, joins the Hudson just downstream of Rockwell Falls. The flow in the lower Sacandaga is controlled by hydroelectric dams, and Stewarts Bridge Dam, the farthest downstream on the river, is located 3.0 mi. (4.8 km) from the confluence with the Hudson River. Water stored in these deep-reservoirs remains relatively warm ($\approx 1°C$) throughout the winter. Below the dam the Sacandaga is steeply sloping, shallow, and generally ice free.

The ice management strategy that we are developing involves 1) minimizing the supply of ice coming from upstream, 2) increasing the water temperature of the Hudson, and 3) increasing the velocity and energy gradient of the flow in the study reach. An ice boom will be installed immediately upstream of Rockwell Falls in a pooled reach of the Hudson that typically remains open throughout the winter. The boom will provide an obstruction that, together with the reduced energy gradient in the pool, will act to retain ice that is moving downstream. If the boom is successful, an ice cover will become established, reducing the heat loss from the river and the quantity of ice entering the study reach.

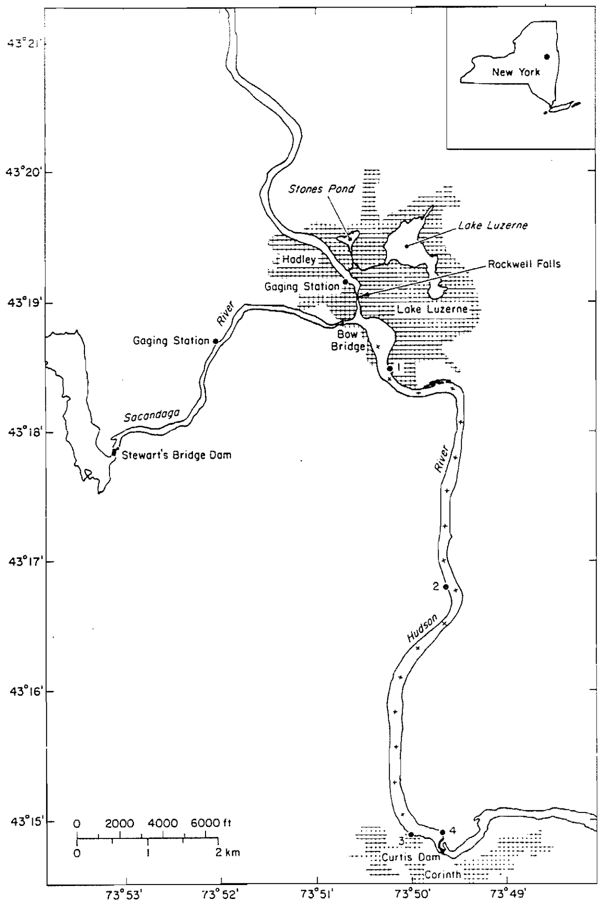

Figure 1: Location map of Hudson River-Sacandaga River study area.

During relatively mild periods, warm water released during normal operation of Stewarts Bridge Dam successfully inhibits the growth of ice in the study reach. In order to optimize this ice control technique, the quantity of flow released should vary with the release temperature, the air temperature and the mixing behavior of the two rivers. As climatic conditions become colder or the quantity of water in storage is depleted, ice control with warm water is no longer possible.

Hydroelectric peak power production can produce abrupt changes in river stage and velocity that delay or prevent the formation of a river ice cover, or cause the breakup of an existing cover. Spring floods, surges generated during breakup and hydroelectric power releases are river waves. In each instance, these long-period, shallow-water waves are a consequence of unsteady flow. Without understanding the dominant physical processes, sufficient observational evidence exists to conclude that river waves are usually an integral part of ice cover breakup. A potential ice management technique for the Hudson River would use river waves generated at Stewarts Bridge Dam to cause the breakup of an ice cover or jam. If this technique is effective, ice cover formation and then removal with scheduled releases would require significantly less water than attempting to continuously melt the ice. However, the quantity of flow, duration, and timing of releases for maximum effectiveness must be determined.

Hudson River Winter Experiment - 1985

We performed our initial set of winter field experiments on 17 January and 9-14 February 1985, focusing on ice cover breakup induced by river waves. Four stations along the Hudson were established (Fig. 1) for ground observations. River stage relative to existing benchmarks was recorded at intervals ranging between 30 s and 5 min., depending upon the rate of change. A log was kept at each station of river and ice cover response. Water temperature was measured periodically during the winter at the gaging station on

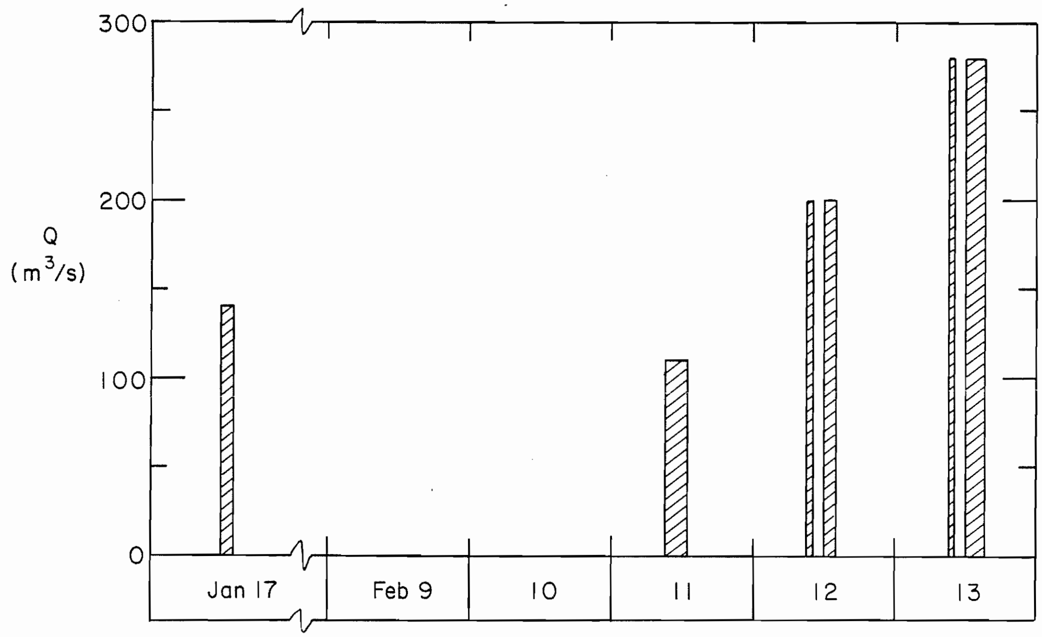

Figure 2: Stewarts Bridge Dam releases, 1985 winter field experiments.

each river (Fig. 1). Average water temperature during the period of 7-16 January was 0.0°C in the Hudson River and 0.9°C in the Sacandaga River. Water temperatures measured between 18 January and 8 February were not significantly different, averaging 0.0°C in the Hudson and 0.8°C in the Sacandaga. Relatively high flows from Stewarts Bridge Dam and generally mild air temperatures greatly inhibited ice growth in the study reach of the Hudson during this particular winter.

The release hydrograph from Stewarts Bridge Dam for the testing periods is given in Figure 2. Base flow measured at the Sacandaga River gage was 1.0 m^3/s during each test. The Hudson River base flow at the Hadley gage averaged 70 m^3/s on 17 January and 40 m^3/s during the February test. The single-pulse January test was conducted with most of the study reach ice-free. However, the river was covered with ice approximately 5 cm thick for a significant distance near station 2. When the 140-m^3/s release arrived at this location, audible fracturing of the cover occurred. The accelerating flow increased the drag on the cover, forcing the entire ice sheet downstream. Shear ridges formed near the banks and irregular pressure ridges formed at many locations in the cover. The fractured ice sheet became unstable, beginning in the center of the channel at the upstream edge of the sheet, and breakup progressed rapidly downstream. The breakup ceased as the stage peaked, and the ice remained motionless throughout the recession.

During the initial two days of the February test all flow releases from the Stewarts Bridge Dam were withheld, allowing a 5- to 15-cm-thick ice cover to become established over the entire study reach of the Hudson River. Over the next several days a series of abrupt releases of varying magnitude, duration and spacing were made from the dam. Aerial black and white photographs (1:5000 scale), taken with a 70-mm camera in an Enviropod mount attached to the underside of the aircraft, were used with supplementary ground observations to monitor ice conditions throughout the February test. Ice type provides information about the ice formation processes at different locations in the reach. Table 1 summarizes the notation used in Figures 3-7 that identifies the various observed ice types.

Ice conditions on 11 February, prior to the arrival of the initial river wave, are shown in Figure 3. The ice in the shallow embayment upstream of station 1, with a thickness greater than 0.3 m, was older and heavier (H) than that found in the remainder of the reach. The embayment is the first place in the reach that forms an ice cover in winter. The next oldest ice, indicated by the presence of a snow cover, was shorefast (SF) ice

Table 1. Ice conditions observed on aerial photographs and during ground surveys.

B	Black ice
BF	Black ice, fragments visible
F	Fragmented ice
FL	Floes
FP	Frazil pans
H	Heavy ice
O	Open water
SF	Shorefast ice
W	White ice
WF	White ice, fragments visible

opposite the embayment and immediately upstream of the right-angle bend in the river. Near station 1 the ice cover was formed from frazil ice pans (FP) that originated upstream and progressed downstream until they became jammed and then frozen into a cover. Immediately downstream of this jam was a reach of black ice (B). Next downstream was a 4.9-km reach with a cover composed largely of black ice fragments (BF) that had refrozen, interrupted near station 2 by a short reach of refrozen white ice fragments (WF). The white enhancement of long linear fractures evident in Figure 3 might indicate frost formation on the ice near a source of water vapor. Finally, the 1.7-km reach upstream of station 3 was covered by white ice (W). This reach apparently contained shorefast ice that entirely bridged the river (Fig. 4). Entrapped air bubbles give the ice a white appearance and indicate rapid freezing, consistent with the low air temperatures recorded prior to and during the zero-release days of the test.

During shutoff of the Stewarts Bridge Dam, a 3- to 5-cm-thick cover of black ice formed on a pooled reach of the Sacandaga River at the gaging station. This ice cover was completely destroyed within minutes after the arrival of the 110-m^3/s river wave on 11 February. Except for the compressed time scale the sequence of events during the breakup parallels those observed at several locations on the Hudson. When the front of the wave arrived the cover began to rise and ice fracturing was audible. Before much change in water surface elevation occurred the entire ice sheet was forced downstream, because of the rapidly increasing shear force, causing pileups of broken ice at rock outcrops and along the banks. Concurrently, the river wave cut an opening through the ice sheet from upstream to downstream near mid-channel. Once the mid-channel was open, large pieces of ice near the banks followed the cross-channel water surface gradient and flowed toward the centerline, and breakup was complete. The wave was next observed about 2 km downstream on the Sacandaga where ice cover fragments of only a few centimeters in size remained.

This 4-hr, 110-m^3/s river wave had a far less dramatic effect on the ice cover in the Hudson River. The initial 1- to 2-hr period following wave arrival produced continuous sound, caused by fracturing of the ice cover. Water fountained up through holes cored in the ice and through shoreline cracks, indicating a pressure flow. Apparent shoreline leads formed near the banks, but the bond between the ice sheet and the ice frozen to the bank remained unbroken. During the rising part of the wave, the fixed ends gave the cover a bowed appearance across the river. The audible ice fracturing and pressurized flow were not observed during subsequent waves later in the test, but the bonding of the ice sheet to the bank persisted in some locations. The only visible changes in the ice sheet resulting from this initial wave were a slight erosion of the upstream FP ice edge and open water pockets in the black ice downstream of station 1 (Fig. 4).

On 12 and 13 February equal discharge releases of different durations were spaced 2 hr apart. The undisturbed river stages measured prior to the first release were lower than the stages at the time of the second release. The cumulative effect on the ice cover of 1-hr and 2-hr releases of 200 m^3/s on 12 February is shown in Figure 5. Except for ice fragments that had jammed and for the shorefast ice, open water extended to about 1.5 km above station 2. The ice moved during the rising part of each wave, with motion ceasing as the wave peaked and all through the recession. The breakup followed the same sequence observed on the Sacandaga River. The high velocity part of the cross section cleared

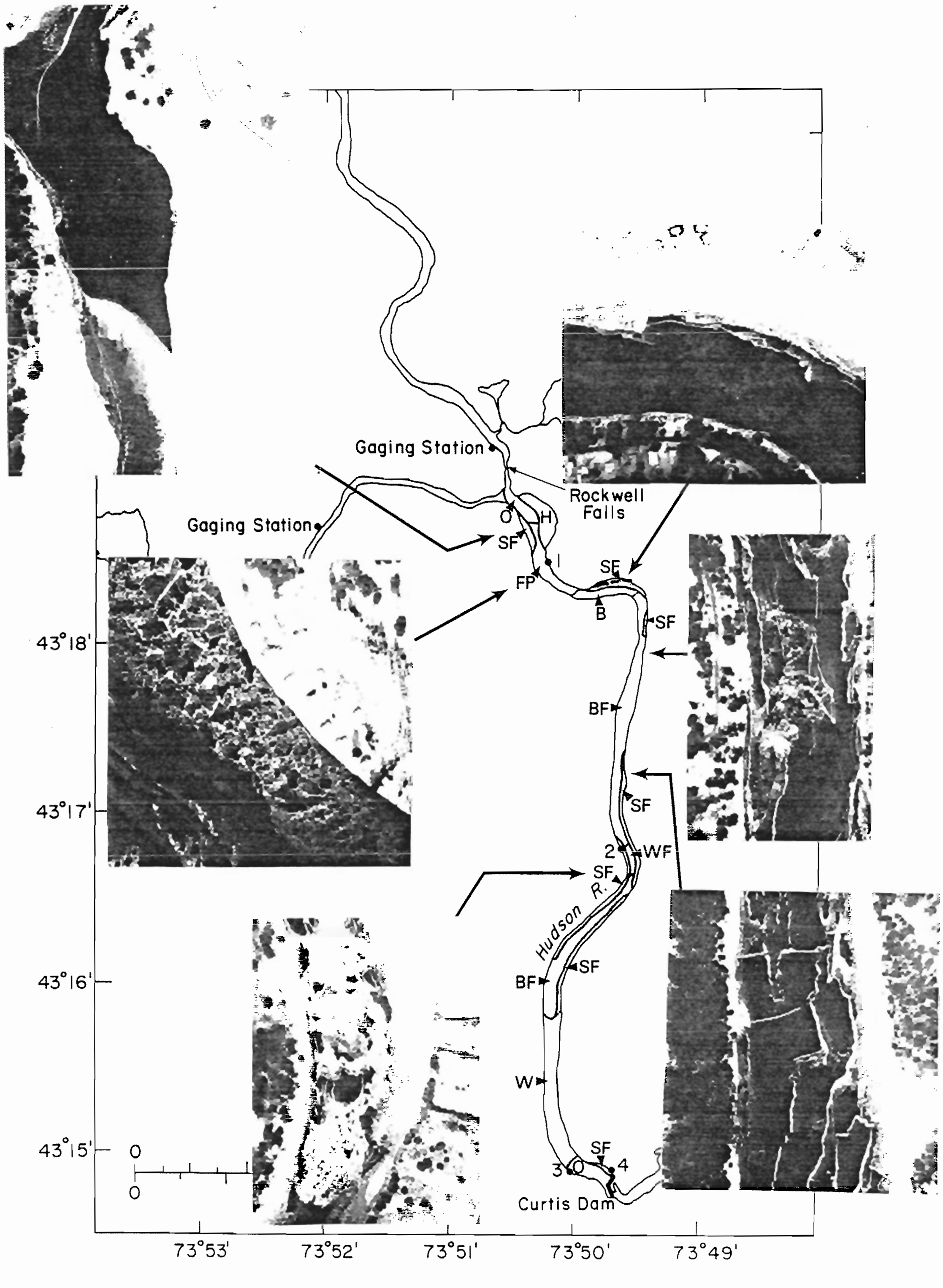

Figure 3: Ice conditions before release on 11 February.

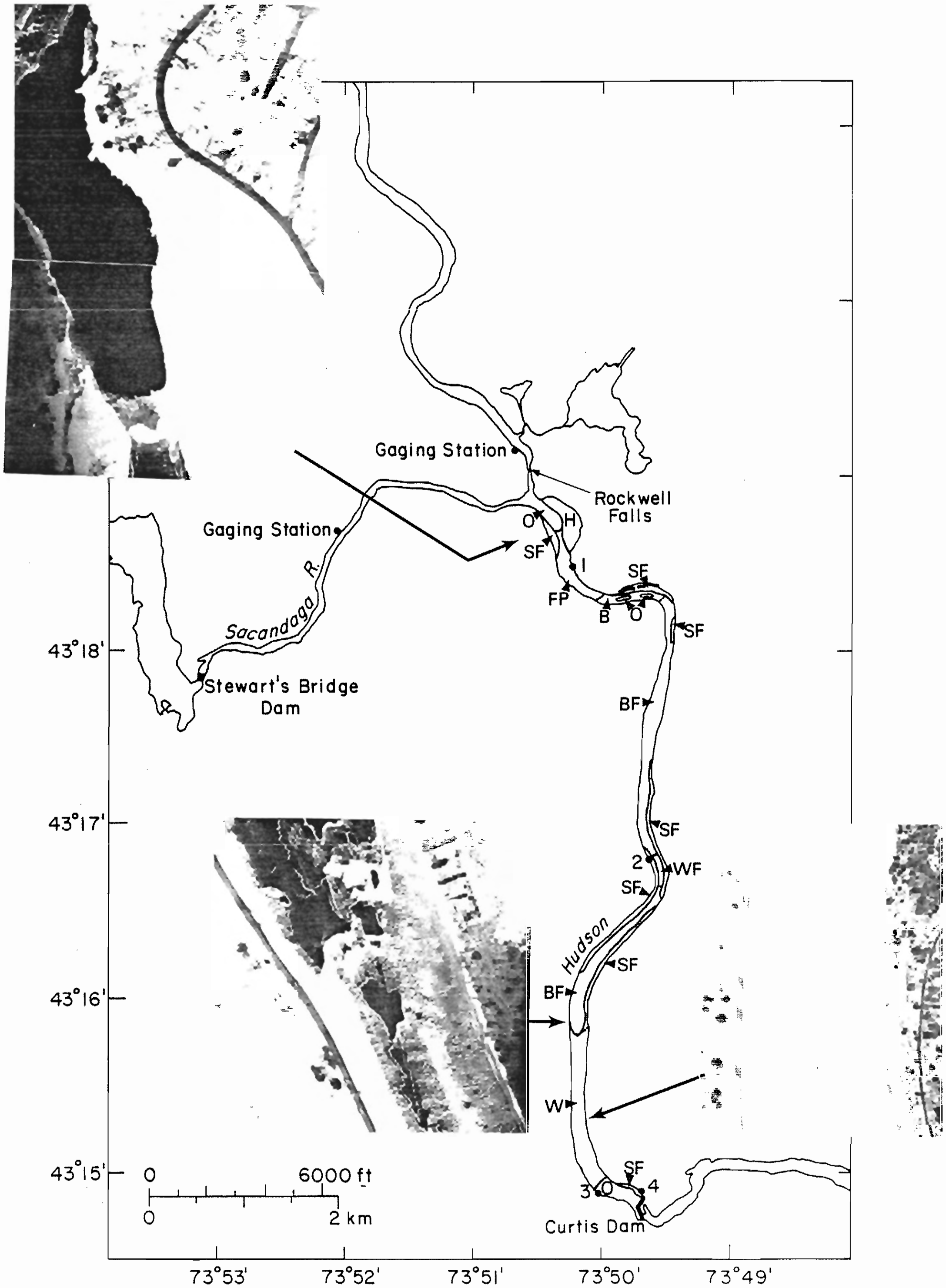

Figure 4: Ice conditions after release on 11 February.

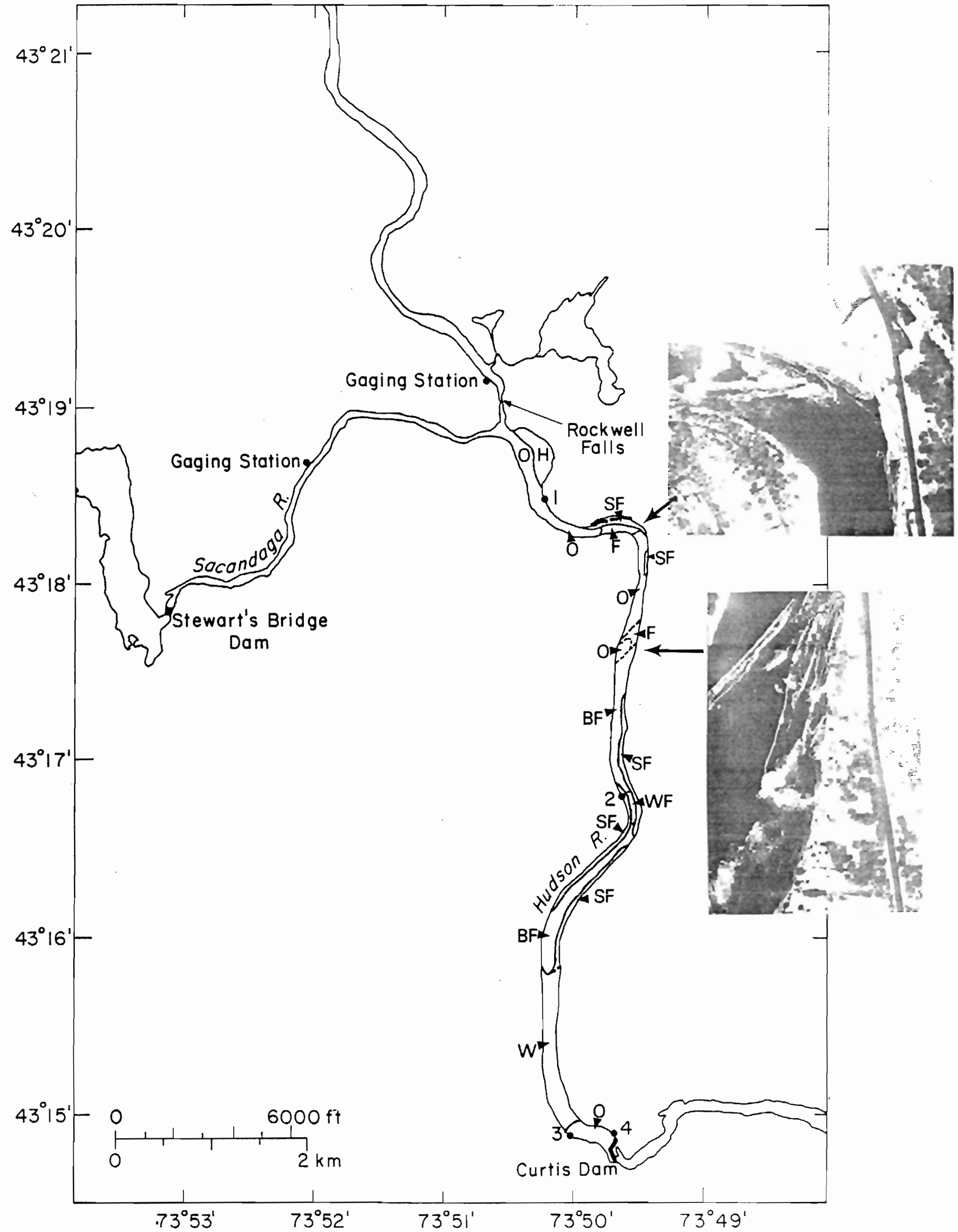

Figure 5: Ice conditions after releases on 12 February 12.

first, followed by movement of large ice floes directly away from shore toward the open water and then downstream.

Figure 6 presents the Hudson River ice conditions on 13 February during the rising hydrograph of the second wave, a 3-hr release of 280 m^3/s. The long piece of shore ice from the left bank upstream of station 2 has broken into two primary floes. Small floes calving from these larger floes and from remaining shorefast ice attest to the highly fractured state of the ice. The large floes that jammed downstream of station 2 were soon

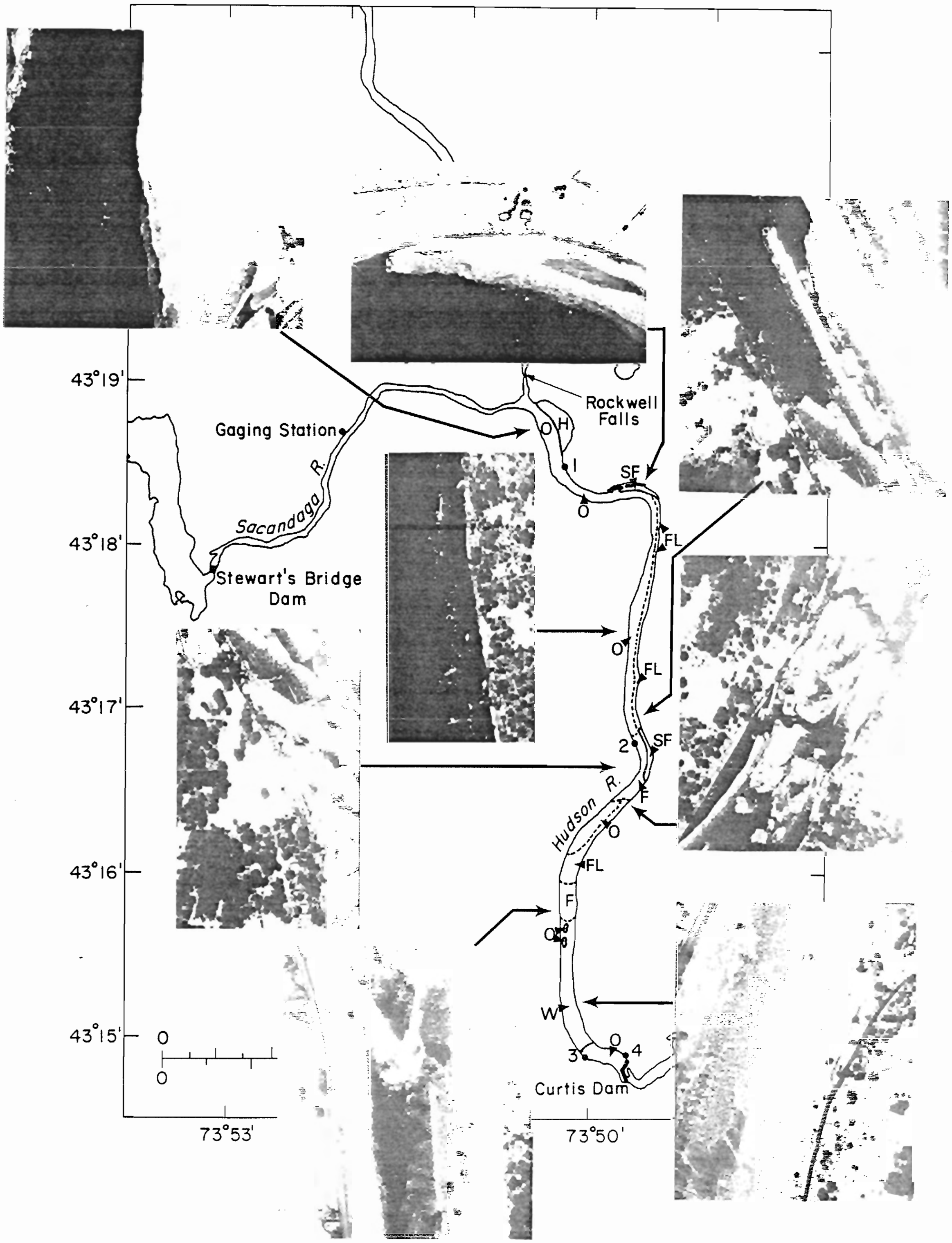

Figure 6: Ice conditions during the second release on 13 February.

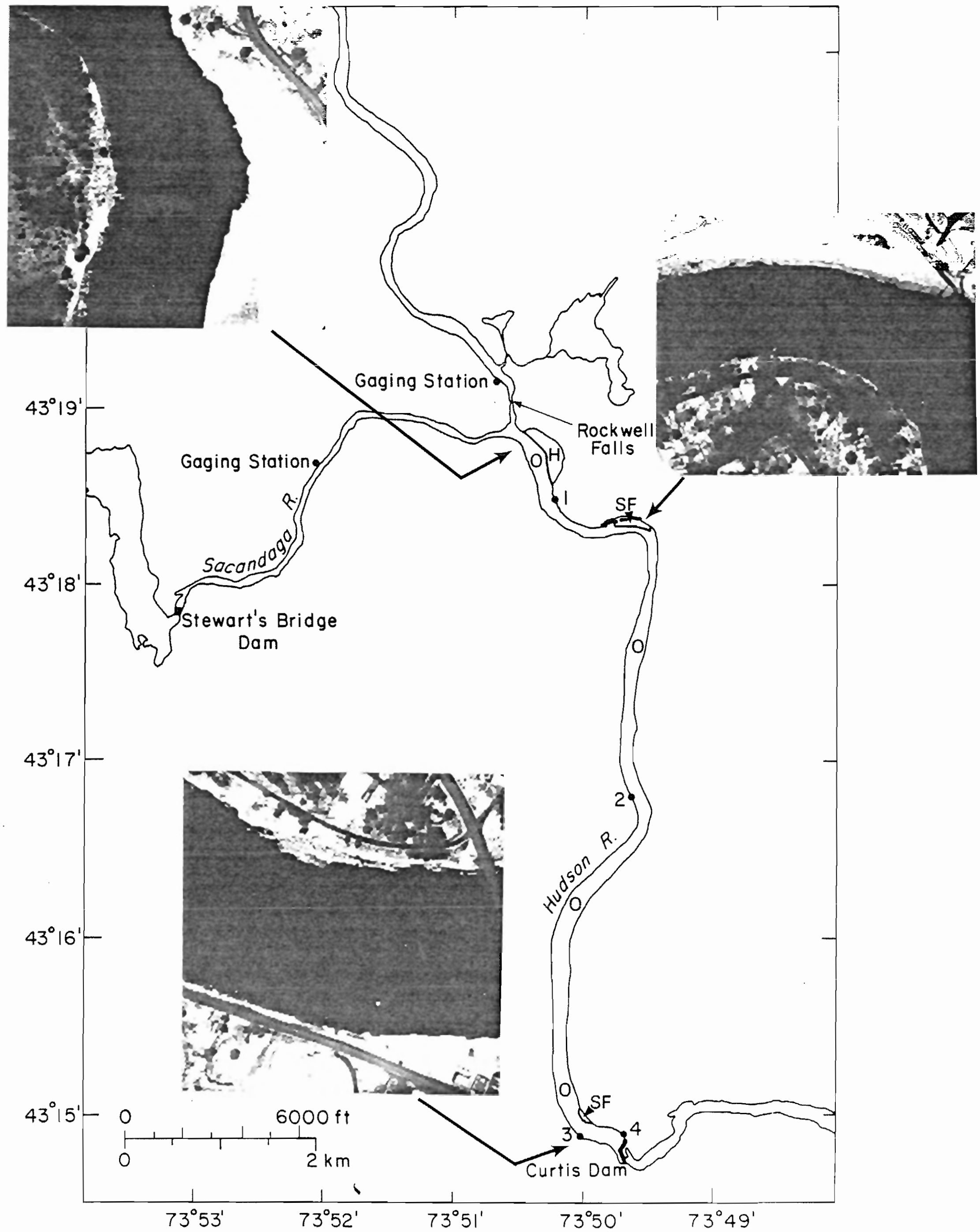

Figure 7: Ice conditions at the conclusion of the test on 14 February.

dislodged. The white ice upstream of station 3 remained intact except at its leading edge. The wedge-shaped breaking front located in the high velocity part of the channel had previously been observed at several upstream locations. When the wave peaked, most ice fragments had joined the ice jam at the head of the white ice, and the progressive upstream to downstream movement of the breakup halted.

A final 2-hr release of 280 m^3/s on 14 February caused the failure of the ice jam and remaining ice cover. Figure 7 presents the ice conditions at the conclusion of the test.

The only significant floes that remained in the study reach were in the embayment and at the right angle bend in the river. The embayment is shallow and protected from the primary flow, and we suspect that similar conditions exist at the bend.

River stage data during the test were collected at the Sacandaga and Hudson River gaging stations and at stations 1-4 (Fig. 1), and results are displayed in Tables 2-5. Wave celerity (speed) in the Sacandaga River increased with discharge (Table 2), and to a lesser extent with the quantity of base flow present in the river. The trend in celerity with discharge is also present in the Hudson River data, except that the 140-m^3/s release had a higher velocity than the larger waves. Greater flow resistance resulting from the presence of the ice cover probably slowed the February test waves enough to account for this discrepancy. We will confirm the effect of an ice cover upon river wave celerity when ice-free summer test results are obtained. An important point to note from these data is the dramatic increase in Hudson River wave celerity relative to that in the Sacandaga River.

The data presented in Table 3 are average rates of stage increase over the initial 12 min. after wave arrival at a given location. Wave front steepness is the change in river stage per unit length of channel. The wave front steepness data in Table 4 were computed by dividing the rate of stage increase at a given location by the wave celerity. At each measurement location, wave front steepness generally increased with greater discharge from Stewarts Bridge Dam and with lower initial river stage. The exception is the lesser steepness of the 140-m^3/s wave at station 4 relative to that of the 110-m^3/s wave. As each wave traveled from the Sacandaga into the Hudson, wave front steepness decreased dramatically. The waves with larger peak discharge had somewhat steeper fronts in the Sacandaga and maintained a significantly greater percentage of that steepness after entering the Hudson. Comparing the data from stations 1 and 4 indicates that front steepness of each wave diminished at nearly a constant rate through the study reach. The 140-m^3/s release, made during largely ice free conditions, exhibited a greater rate of steepness attenuation.

Table 2. Wave Celerity (m/s).

Discharge (m^3/s)	Duration (hr)	Sacandaga River	Hudson River
110	4	1.36	4.57
140	2	1.44	5.39
200	1	1.64	5.03
200	2	1.80	5.03
280	1	1.91	5.21
280	3	2.01	5.21

Table 3. Rate of Stage Increase at Wave Arrival (m/min.).

Discharge (m^3/s)	Duration (hr)	Sacandaga Gage	Hudson Station 1	Hudson Station 4
110	4	0.069	0.011	0.0065
140	2	0.077	0.015	0.0067
200	1	0.091	0.020	0.010
200	2	0.086	0.016	0.0090
280	1	--	0.029	0.018
280	3	0.12	0.025	0.014

Table 4. Wave Front Steepness (m/m).

Discharge (m^3/s)	Duration (hr)	Sacandaga Gage ($x10^3$)	Hudson Station 1 ($x10^3$)	Hudson Station 4 ($x10^3$)	Hudson 1 / Sac. Gage	Hudson 4 / Hudson 1
110	4	0.85	0.041	0.024	0.048	0.59
140	2	0.89	0.048	0.021	0.054	0.44
200	1	0.93	0.066	0.034	0.071	0.52
200	2	0.80	0.052	0.030	0.065	0.58
280	1	--	0.092	0.057	--	0.62
280	3	1.0	0.079	0.045	0.079	0.57

Table 5. Wave Amplitude (m).

Discharge (m^2/s)	Duration (hr)	Sacandaga Gage	Hudson Station 1	Hudson Station 4
110	4	1.40	0.60	0.39
140	2	1.61	0.48	0.30
200	1	1.84	0.41	0.26
200	2	1.75	0.59	0.37
280	1	--	0.59	0.39
280	3	2.16	0.89	0.57

The wave amplitude data presented in Table 5 represent the change in water surface elevation from immediately prior to wave arrival to the peak elevation at each measurement location. Wave amplitude consistently diminished with distance downstream, and was greatly affected by the duration of the release. Longer duration releases produced larger wave amplitudes on the Hudson than those of shorter duration and equal discharge. The 4-hr release of 110 m^3/s produced greater wave amplitudes than all larger releases of shorter duration except those of 280 m^3/s. There was no relationship between wave amplitude and ice-breaking effectiveness. Sufficient stage increase to create additional channel width did not occur with any wave, and was not a factor in the breakup.

The waves with the largest peak discharges had the greatest front steepnesses and were most effective at breaking the ice. These waves did not create appreciably greater water depths, but did cause much higher flow velocities than waves of smaller discharge. The wave peaks at stations 1-4 were never flat, indicating that releases with durations greater than 4 hr are needed to establish steady flow in the study reach. The Curtis Dam powerhouse was not operating during the tests and the dam behaved as an overflow weir. As a result, the rate of river recession was governed by the water level.

Comparison of River Ice Breakup

The primary limitation of the 1985 Hudson River experiments was the relatively thin ice cover. In addition, some melting of the ice occurred on 12-13 February because of mild air temperatures and a mixed river temperature of 0.6°C during the releases. The processes leading to breakup of this ice cover may be different from those of thicker ice that may occur on the Hudson in other years. Therefore, we will compare our observations with those made of other rivers having thick ice covers.

Andres and Doyle [1984] discussed breakup on the Athabasca River at Fort McMurray, Alberta, for the years 1977-1979. The ice cover was approximately 1 m thick at the time of breakup each year, which coincided with the passage of large-amplitude river waves. They reported that steep, rapidly flowing reaches were the first to experience breakup at relatively small increases in discharge. The broken ice accumulated in low velocity reaches downstream and formed jams. As the discharge continued to increase, the jams moved downstream, combined, and alternately moved and reformed. The ice run stalled and a massive jam formed downstream of Fort McMurray, where the river becomes wider and flatter. The ice floes in this jam decreased in size in the upstream direction. Whether the jam failed rapidly or deteriorated slowly in a given year depended upon the rate of discharge increase.

Anderson [1982] and Prowse [1984] studied the 1982 and 1983 breakups, respectively, of the Liard River in the vicinity of Fort Simpson, N.W.T. The observations of events leading to breakup were similar in both years, and we will summarize the more detailed description of Prowse [1984]. Several measurements taken prior to breakup revealed that the river ice cover was solid and about 1 m thick. The ice breakup front progressed downstream with a spring river wave. It moved in a series of surges and stalls corresponding to the failure and reformation of ice jams. There was a rapid transition from larger to smaller ice floes with distance upstream of the breaking front. Breakup near the center of the channel in steep, high-velocity reaches and formation of large apparent shore leads occurred far ahead of the primary wave and breakup front. Closer inspection of the leads revealed that water was overlying intact ice and the ice sheet remained bonded to the shore. The rate of ice clearance from the Liard River varied between 65 km per day from Fort Nelson to Fort Liard, to about 2.5 km per day near the confluence with the Mackenzie. Within about 20 km of the river mouth the bed slope decreases by almost an order of magnitude (from 0.00039 to 0.00005) and the width and cross-sectional area expand.

The descriptions of breakup on the Athabasca and Liard Rivers are very similar, and parallel our observations of the Sacandaga-Hudson breakup. The early breakups in steep reaches of the Athabasca and Liard at minimal increases in discharge are analogous to the early and abrupt Sacandaga River breakup. The increasing discharge in the spring created waves on the Athabasca and Liard that caused ice breakups from upstream to downstream in a sequence of surges and stalls. In the Hudson a series of small-amplitude waves successively moved the breakup farther downstream. The wave front steepness controlled failure of the massive jam at Fort McMurray. Similarly, increasing wave front steepness on the Hudson was directly related to ice breakup effectiveness. Smaller ice floes located farther upstream in the reach of fragmented ice were typical on the Athabasca and Liard. Observations of breakup at Hudson River stations 1 and 2 indicated that ice floe sizes were generally larger when local ice cleared from a reach, but ice floes traveling from upstream were smaller. Our observations indicate that breakup of large floes having preexisting cracks is a rapid process, and significant size reductions can occur in distances of only a few kilometers. The report of early formation of apparent shore leads on the Liard is analogous to the Hudson River ice cover response. Actual shore leads typically developed much later, often near the time of local ice breakup. The rate of ice cover clearance from the Liard River was related to the channel slope, slowing as the slope decreased. The tendency for jamming when the slope becomes sufficiently small is similar on the Athabasca, Liard and Hudson Rivers.

The strength of a river ice cover is related to its thickness. However, comparing the Sacandaga-Hudson breakup observations with those of the Athabasca and Liard Rivers indicates that the basic physical processes are independent of ice thickness. Our Hudson River studies have confirmed previous observations of fundamental river wave involvement in ice breakup.

Analysis of River Ice Breakup

Development of a relationship between the physical parameters of a river wave and ice breaking capability would further the understanding of breakup and provide a theoretical basis for using waves to manage ice. Wave front steepness is closely related to the ice breaking capability of a river wave. Billfalk [1982] analyzed the initial breakup of a solid river ice cover resulting from rapid water level changes associated with the passage of a river wave. He reasoned that water level change induces a bending moment in the ice, which causes fracturing of the cover when it exceeds the flexural strength of the ice. The critical slope of the wave front was defined as the minimum slope required for bending failure. The theory of beams on an elastic foundation was applied to analyze a semi-infinite strip of ice having unit width and oriented along the channel. The assumptions made were that the wave front is linear, the channel is either wide or the ice cover has open cracks along the banks, and the ice undergoes elastic deformation and brittle fracture. Buoyancy forces were considered important, but drag and inertia forces acting on the ice were ignored.

The critical wave front steepness for bending failure computed for the Hudson River, with an assumed ice thickness of 10 cm, is 0.015. A critical steepness of 0.020 is found for the Sacandaga River with an assumed ice thickness of 3 cm. However, because simplifications used in the computation do not hold for thin ice, the required wave steepnesses for bending failure are actually greater. Comparing critical and measured longitudinal wave steepnesses (Table 4), we find that bending failure of the ice should not occur in either river. Longitudinal bending failure of the ice was not observed at any location during the test.

The acoustic emission that accompanied the initial wave on the Hudson indicated extensive fracturing of the ice sheet. A mechanism responsible for much of this fracturing was probably cross-channel bending. The hinge cracks did not release when the wave arrived, causing a 1-m elevation difference between the near shore and midchannel parts of the sheet. A typical transverse slope of 0.020 developed, of the same order of magnitude as required to cause bending failure of the ice. However, the ice sheet was not broken into obvious fragments, and the apparent breakup had not begun. Some wave front steepness was needed to produce bending and fracture the ice but the wave parameter associated with front steepness that was responsible for the breakup is not resolved.

We propose the energy slope as the driving force in the breakup of river ice covers and ice jams. A larger energy slope corresponds to greater ice breaking capability. Resistance to breakup is provided by ice cover strength and the stability of the floes in the fractured cover. The momentum equation for unsteady flow in a river can be written as [Stoker, 1957],

$$S_f = S_o - \frac{\partial y}{\partial x} - \frac{1}{g}\left(\frac{\partial v}{\partial t} + v\,\frac{\partial v}{\partial x}\right) \tag{1}$$

where S_f is the energy slope, S_o is the river bed slope, y is the depth, v is velocity of the flow, g is acceleration due to gravity, x is longitudinal distance, and t is time. When a river is steeply sloping the energy slope is also large, and steep river reaches are the first to experience breakup. Ice runs tend to stall against an existing cover and form jams where river slope decreases. River waves cause the energy slope to differ from the bed slope. At any location along the river, S_f is larger during the rising part of a wave and smaller during recession than for corresponding steady flow conditions. The relative size of the increase or decrease in S_f depends upon river and wave properties.

Properties of river waves such as wavelength, celerity, front steepness,wave height and rate of attenuation can vary greatly. Ferrick [1985] found that river waves can be characterized by the following dimensionless scaling parameters:

$$F_I = \frac{2\,C_r}{C_*^2}\left(\frac{k\Delta x}{y_o}\right) \tag{2}$$

$$C_r = \frac{v_o}{c_m} \tag{3}$$

$$F_c = F_I\,C_r \tag{4}$$

$$D = \frac{C_*^2}{2\,F_o^2\left(\frac{k\Delta x}{y_o}\right)} \tag{5}$$

Table 6. River Wave Parameters.

River	k	v_o (m/s)	y_o (m)	c_m (m/s)	Δx (m)	C_*^2	F_I	F_c	C_r	D	Wave Type
Athabasca	2	2.0	4.8	4.0	5500	71	32.	16.	0.50	0.18	Bulk
Hudson	1	0.34	3.5	5.4	4500	97	1.7	0.11	0.063	--	Dynamic
	2	0.43	3.7	5.2	4500	102	3.9	0.33	0.083	--	Dynamic
Liard	2	1.0	4.0	1.5	155,000	80	1300	860	0.67	0.020	Bulk
Sacandaga	1	1.0	0.85	1.4	2400	27	150.	110.	0.71	0.040	Bulk

where C_* is the dimensionless Chezy conveyance coefficient, Δx is the lesser of the half-wavelength or half reach length, y_o is the mean depth, v_o is the mean velocity, F_o is the Froude number evaluated at mean depth and velocity, c_m is measured wave celerity, and k is a parameter accounting for the presence (= 2) or absence (= 1) of an ice cover. The parameters F_I and F_c indicate the relative importance of friction and inertia in the flow momentum balance. Large values of F_I and F_c indicate a large energy slope and greater capability to break ice. The primary river wave types are friction-dominated bulk waves (F_I, $F_c > 10$), dynamic waves ($10 > F_I$, $F_c > 0.1$), and inertia-dominated gravity waves (F_I, $F_c < 0.1$). The parameter D characterizes the rate of attenuation of bulk waves.

The scaling parameters characterize the average behavior of a river wave in a given reach. Values of the scaling parameters computed for a particular location and time will vary from these averages because wave and channel characteristics change with location and time. However, the parameters are useful for interpreting local wave behavior if the changes in physical characteristics are known. We calculated average scaling parameters for the Hudson, Sacandaga, Athabasca and Liard Rivers and present them in Table 6.

An ice jam release initiated a river wave that caused breakup of competent ice cover in an 11-km reach of the Athabasca River near Fort McMurray in 1979. The data characterizing the wave [Ferrick, 1984] yield large values of F_I and F_c (Table 6), indicating that a large energy gradient and ice breaking capability were associated with the wave front. The flow parameters given for the Liard River in Table 6 are estimates that were based upon data presented by Prowse [1984]. The wave celerity value given was the average speed of the breaking front over the 311 km measured upstream from the river mouth. Movement of the breaking front at celerities of up to 5 m/s were measured in relatively short subreaches, but additional data are needed to interpret this local behavior. The wavelength and reach length considered for the Liard are much greater than for the other rivers. At a comparable length scale Δx, the F_I and F_c values representing the Liard are not greatly different than those of the Athabasca. Again, a large energy slope and ice breaking capability are indicated.

Waves in the Sacandaga River behave as bulk waves with characteristically low celerities. The small D value indicates that an abrupt release from the dam will retain a sharp front as it approaches the confluence with the Hudson. The F_I value of the 140 m^3/s release is two orders of magnitude smaller for the Hudson than is the corresponding value for the Sacandaga, indicating a change from bulk to dynamic behavior as the wave enters the Hudson. This change in wave type is evidenced in the data as a drastic reduction in front steepness and increase in celerity as each wave enters the Hudson. The second Hudson River calculation is for an ice-covered case with a mean discharge of 180 m^3/s. The presence of the ice cover and larger Courant number C_r resulting from the increase in discharge cause F_I to more than double. Our observations of greater ice breaking effectiveness at larger discharge and smaller initial depth correspond to maximum F_I.

Conclusions

River waves provide an effective tool for ice management in the study reach of the Hudson. Breakup and movement of the ice occurred on the wave fronts, where unsteady flow effects increase the energy gradient at the breaking front relative to steady or diminishing flow conditions. Wave fronts having the largest energy gradients caused the farthest downstream movement of the breaking front. Based on these observations we propose the energy gradient of the flow as the parameter that quantifies the driving force for ice cover breakup. However, we recognize that additional evidence and study are needed to support this theory.

The range of Hudson River ice thicknesses that can be effectively managed with river waves is not yet well defined. The ice breaking capability of these waves is constrained by the maximum energy gradient that can be developed at each point in the reach. The large channel capacity and flat bed slope of the Hudson, and the limited flow capacity of the lower Sacandaga determine this upper bound. The minimal ice thickness with reduced resistance to breakup that was present during the 1985 field experiments adds a second factor to the question of wave effectiveness during heavier ice conditions. Because of these uncertainties, prudent ice management would use waves as needed throughout the winter to prevent thick accumulations. Prior to the spring ice run, waves would be used to remove an existing ice cover, minimizing the potential for jamming.

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