Effects of Permafrost on Stream Channel Behavior in Arctic Alaska

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By KEVIN M. SCOTT

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By Kevin M. Scott

ABSTRACT

Sites with drainage areas ranging from 88 to 12,200 km² were monitored on five streams in northern Alaska during the breakup in 1976 to determine (1) the effects of frozen bed and bank material on channel behavior, and (2) the importance of the annual breakup flood in forming the channels of arctic streams.

The thawing and concomitant erosion of channels varied with changes in bed-material size, channel pattern, drainage area, and climate. The response of channels to breakup flooding ranged from total permafrost control of channel processes, including both bed scour and lateral erosion, to only brief restriction of channel behavior early in the rise of the flooding. The watershed characteristic that appears to explain much of this variation is size of drainage area.

Similar variation of channel response with these factors is believed to be responsible for diametrically opposite results reported from recent studies of the two problems posed above. Permafrost has been cited both as the cause of extreme stability and as the cause of unusual instability in arctic streams as compared to those elsewhere. That permafrost simplifies the study of arctic stream channels through its domination of the effects of other variables has been a common assumption. As a consequence, however, the generalizations based on a single stream or on similar streams have led to a spectrum of inconsistent results. This spectrum of previous results now appears potentially consistent.

Comparisons of absolute rates of lateral erosion are not feasible, but it is likely that the net effect of the permafrost environment is to create greater channel stability than is found in unregulated streams of similar size in nonpermafrost environments. Combinations of factors, particularly those that encourage high rates of thermo-erosional niching, can nevertheless cause rates of erosion that dictate caution in engineering design.

INTRODUCTION

Many studies of arctic streams have been made since 1970 in response to an increased concern about development, particularly of oil and gas resources, in regions of continuous permafrost. This expansion of interest has occurred simultaneously in Alaska, Canada, and the U.S.S.R. It has focused especially on the relationships of permafrost to the hydrology of streams and to the behavior of stream channels. Both these aspects are fruitful areas of geomorphic research with applications to the engineering of structures in flood plains and stream channels. It is the latter aspect, however—how and when channels in the permafrost environment respond to flow—to which this investigation is confined.

Two questions have been of recurring interest in studies of the behavior of arctic streams: (1) What is the effect of permafrost on rates of bank erosion—does permafrost inhibit lateral migration, or do the streams of permafrost areas have higher lateral erosion rates than might be expected for channels that contain effective flow less than half the year? And, (2) what is the role of the spring breakup flooding, compared to flow during the rest of the runoff period, in determining the channel characteristics; that is, what is the annual timing of the channel-forming processes?

In addressing the first question several recent workers have agreed with Leffingwell's (1919, p. 172) conclusion that lateral erosion is generally retarded by frozen ground. Cooper and Hollingshead (1973, p. 276) note that permafrost stabilizes bank material that, if unfrozen, would offer little resistance to erosion. McDonald and Lewis (1973, p. 2) believe that permafrost retards erosion over short time spans. It also has been suggested that, because the annual floods of some Alaskan streams may occur while channels are frozen, the channel size is anomalously small in relation to flood discharges (see comment by C. R. Neill in Kane and Slaughter, 1972, p. 540).

Walker and Arnborg (1966, p. 171) and Ritchie and Walker (1974, p. 545), on the other hand, emphasize the role of permafrost in promoting notable lateral erosion (an average of 10 m in one area) by lateral undercutting during breakup of the Colville River in northern Alaska. Walker (1973, p. 73) notes a general belief that lateral erosion is "especially important" in arctic Alaska because of permafrost. High rates of erosion of bedrock permafrost by streams have been attributed to an "ice rind," an easily eroded surficial zone that has been shattered by ground ice (Büdel, 1972, p. 9).

The engineering reports dealing with pipeline construction also manifest differing opinions on the stability of arctic streams. One 1975 unpublished report cites the "extreme" stability of a reach of the Sagavanirktok River in northern Alaska as being related to the presence of permafrost in the flood plain. Another, compiled in 1974 and dealing with rivers on the Arctic Coastal

Plain east of the Sagavanirktok River, notes that such rivers have a "greater tendency . . . for lateral migration . . . than is the case for most rivers in other physiographic regions." The significance of these comments is less in their substance than in the fact that they exist without contradiction in the scientific literature.

Which opinion is valid—are the streams of arctic Alaska extremely stable or do they in fact have greater-than-average rates of lateral migration? More definitive work is plainly needed. An example of the practical reasons for the recent concern over rates of lateral erosion is the placement of pipeline sag points, the sites of downward inflection of the pipe to depths safe from exposure by scour (short-term vertical erosion of bed) or longer term degradation of the streambed. Because deeper burial is costly and involves other problems, the length of deep burial is minimized in compliance with estimates of the probable rate of stream migration at each point of stream-pipeline proximity.

Evidence on the second question—the importance of breakup flooding and of the annual timing of the processes that significantly influence the beds and banks of streams—is just as diverse. Walker and Arnborg (1966, p. 168–171) found that most lateral erosion occurs during or shortly after breakup flooding on the Colville River, as did Outhet (1974, p. 308) for each of five categories of bank shape in the Mackenzie River delta. At the other extreme, Miles (1976, p. 196 and 199) found that breakup on Banks Island in the Canadian Arctic was relatively unimportant in this respect, and that significant change in the channels did not begin until summer storms occurred several weeks later. To complete the possible variations, Abramov (1957, p. 112) reported that bank slumping along the lower Lena River in Siberia was greatest toward the end of summer.

These conclusions suggest that the behavior of arctic streams, in response to a relatively simple hydrologic regime of spring breakup flooding followed by low summer flow with occasional rainstorm runoff events, is far from uniform. Some studies of arctic streams have implied the extension of conclusions representing perhaps a single stream over one season to arctic processes in general. This study is not an exception to this tendency, but it approaches the problem through the study of multiple sites on a group of five streams representing mountain, foothill, and coastal-plain drainage areas. A degree of perspective is added to the observations by flow records of from 5 to 7 years duration on all but one of the five streams. The study periods were May 5 to June 14, 1976, an interval that included the rise and recession of breakup flooding in most of the streams, and May 24 to June 1, 1977.

The study attempts to show the effect of permafrost on streams by concomitant measurement of lateral erosion and of the rate of retreat of the thawed zone in the eroding bank during the period of breakup. Care was taken to define two factors that are believed to be closely related to any observed variation in erosion rates—size of bed and bank material as a cause, and channel pattern as an effect. Although any of several factors can affect channel pattern, Brice (1964, p. 32) regards bank erodibility as the most significant single variable. Bank erodibility in turn depends mainly on the size distribution of the bed and bank material (see Schumm, 1960, p. 17 and 22–23), as well as on the vegetation of the banks (Mackin, 1956; Smith, 1976), and, in the case of cohesive material, on the moisture content of the bank in combination with freezing and thawing (Wolman, 1959, p. 214).

ACKNOWLEDGMENTS

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DESCRIPTION OF STREAMS

The five study watersheds drain the almost treeless desert or semidesert that constitutes much of the north slope of the Brooks Range (fig. 1). The area is one of continuous permafrost in which the active layer, or zone of seasonal thaw, reaches a late-summer thickness of less than 0.3 m in tundra-covered cohesive material and as much as 2 m in unvegetated noncohesive material. Modern glaciers, present only in high mountain valleys above 1500 m altitude, cover less than one percent of the drainage area of the Atigun River—on average the highest and most mountainous of the watersheds studied.

Mean annual precipitation, although not well documented, is probably less than 200 mm over much of the coastal plain and increases with altitude to larger totals near the crest of the Brooks Range. Approximately half of the precipitation falls as snow from September through May, and the resulting meltwater runoff is largely concentrated in the breakup flooding of late May and early June. Summer rains are associated with the semipermanent arctic front that oscillates irregularly between the Brooks Range and the Arctic Ocean. The temperature gradient with latitude may be pronounced and, associated with the effects of coastal fog, can result in delay of breakup in streams draining the coastal areas by several weeks beyond that of streams with their headwaters in the mountains and foothills. The temperature gradient reverses with the season, however; winter temperatures are frequently between −26° and −35°C at the coast and −30° and −40°C inland, and summer temperatures are commonly between 5° and 13°C near the coast and 10° and 18°C inland.

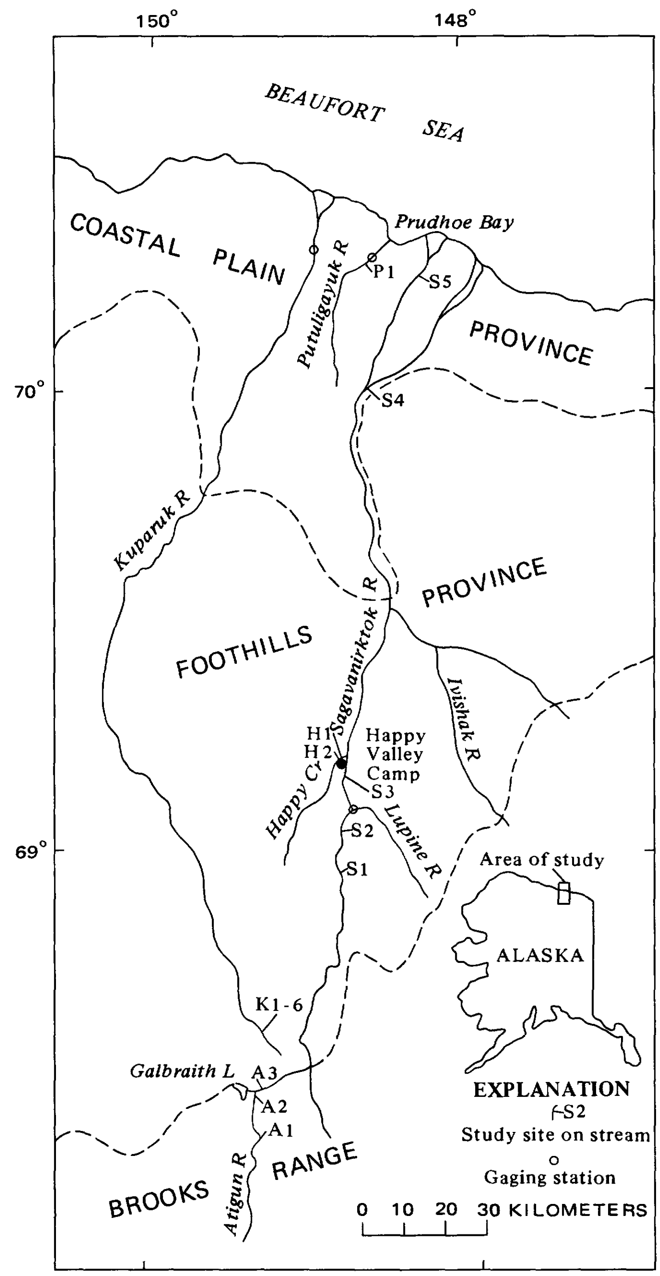

FIGURE 1.—Map showing locations of study sites.

Stream valleys within the Brooks Range were the sites of late Pleistocene mountain glaciers that extended as much as 50 km north of the range front (Hamilton and Porter, 1975, p. 480). Unglaciated uplands and flood plains were profoundly affected by periglacial processes such as solifluction and the formation of patterned ground—processes continuous today but less intensively active. The larger valleys in the Sagavanirktok River headwaters are glacially sculptured, floored with glacial-outwash sediment, and bordered by alluvial fans formed by debris flows from tributary hanging valleys.

The channel patterns vary in a downstream direction with changes in several factors, the most obvious of which is bank erodibility. The Sagavanirktok River, for example, changes from braided to sinous or meandering and back to braided during its course through the foothills. A similar sequence in pattern occurs on the Atigun River and can be ascribed to the same basic situation—a lower gradient in reaches upstream from points of control by bedrock or moraines has resulted in deposition of fine-grained alluvium. The cohesive nature of the fine-grained sediment results in reduced bank erodibility that is accompanied by the development of a sinuous or meandering pattern. A braided pattern occurs downstream as well as upstream from the sections with fine-grained alluvium. In most instances the stream pattern is clearly braided or sinuous-meandering, but in others the pattern appears to be transitional, similar in some respects to the pattern called wandering by Mollard (1973, p. 347 and fig. 4). Channels may be generally braided at lower stages and clearly sinuous or meandering at intermediate and high stages. Channel patterns are described in table 1 according to a scheme (J. C. Brice, written commun., 1977) in which streams with a sinuous pattern have a sinuosity of 1.06 to 1.25, and those with a meandering pattern have a sinuosity between 1.25 and 2.0. Sinuosity is the ratio of length of channel to length of meander-belt axis, or to length of valley axis if no meander belt can be defined (Brice, 1964, p. 25).

The fine-grained alluvium forming the banks of sinuous and meandering channels generally consists of cohesive silt and clay with organic-rich layers and associated soil development in the upper 0.3 to 0.7 m. Gravel stringers are commonly intercalated in the basal part of the silt-clay alluvium, which may be as much as 4 m in thickness, and the unit overlies coarse channel deposits with a texture closely approximating that of the bed material of the stream. The unit may thin laterally, and locally may consist only of a predominantly organic tundra mat. Much of the fine-grained alluvium apparently consists of reworked loess deposited during overbank flooding and stabilized by tundra vegetation.

The most obvious form of lateral erosion in the streams is thermo-erosional niching—the process of undercutting of frozen banks by thawing and erosion at the same time—described from the Colville River delta by Walker and Arnborg (1966, p. 166–167). Banks may be undercut, or "niched," to surprising distances by breakup meltwaters.

All observations indicated that the banks and beds at the study sites were frozen at the start of the runoff period. A possible exception was the downstream part of the Sagavanirktok River (site S5), where measurements could not be obtained and the year-round existence of a shallow aquifer in the stream alluvium has been indicated (Sherman, 1973). The degree to which the Sagavanirktok River channel is underlain by permafrost is controversial, and the observations here of a frozen bed at the start of a breakup at all sites upstream from site S5 do not unequivocally indicate immediately underlying permafrost, although the evidence is strong. The observed frozen bed material, even in the thalweg of the channel, conceivably could represent seasonal frost above an underlying unfrozen zone. As far as stream channel behavior during the breakup period is concerned, the difference is not important. The typical case for at least all but the largest coastal-plain streams in their downstream sections is probably that observed in the Shaviovik River, 35 km east, beneath which winter freezing is complete but permafrost temperatures are significantly higher than at locations away from channels (Brewer, 1958, p. 26).

Streamflow in Alaska north of the Brooks Range is measured at present in three major streams—the Sagavanirktok, Kuparuk, and Putuligayuk Rivers (see measurement sites in fig. 1). Flow records since 1970–71 (fig. 2) indicate that the maximum yearly discharge in these streams usually occurs during the breakup flooding and that the breakup peak has occurred most commonly in early June. This flow pattern is characteristic of a "nival" runoff regime (Church, 1974), in which the spring breakup flood is usually the dominant hyrdrologic event. The term "breakup flood" is used in this discussion as an equivalent of "spring," "nival," or "snowmelt" flood—terms that are likewise in wide use.

The hydrographs of the main runoff periods in figure 2 show the variations in amount and distribution of runoff seasonally, from year to year, and between stations. The available records of flow, chemical analyses, water temperatures, and suspended-sediment loads are reported in the annual series of the U.S. Geological Survey, "Water Resources Data for Alaska." Each of the five study watersheds is discussed below.

TABLE 1.—*The study sites and their characteristics*

[Extent of glaciation mainly after Keller, Morris, and Detterman (1961) and Hamilton and Porter (1975). Where late Pleistocene glaciation is indicated, earlier glaciation may also have occurred]

Stream	Site No.	Drainage area (km^2)	Median diameter of bed material (mm)	Channel slope (m/m)	Channel pattern	Glacial history	Site location
Atigun River	A1	435	22	0.00088	Meandering	Multiple late Pleistocene	18.2 km upstream from pipeline crossing.
Do	A2	600	.41	[1].001	Sinuous	do	1.1 km upstream from pipeline crossing.
Do	A3	760	.38	[1].001	do	do	0.15 km upstream from pipeline crossing.
Kuparuk River	K1–6	132–146	56–77	.0059	Meandering	Pre-late Pleistocene	0.06 km upstream to 2.4 km downstream from pipeline crossing.
Sagavanirktok River.	S1	4,680	80	.0032	do	Single late Pleistocene	25.4 km upstream from Lupine River.
Do	S2	4,830	86	.0023	Sinuous	Possible pre-late Pleistocene	9.3 km upstream from Lupine River.
Do	S3	5,870	45	.0025	Braided	Unglaciated	8.7 km downstream from Lupine River.
Do	S4	12,200	18	.0018	Complexly braided	do	Point where distributaries bifurcate.
Do	S5	Distributary	15	.00053	do	do	17.6 km upstream from coast on west channel.
Happy Creek	H1	88	51	.009	Meandering	do	2.7 km upstream from Sagavanirktok River.
Do	H2	89	35	.01	Sinuous	do	1.9 km upstream from Sagavanirktok River.
Putuligayuk River.	P1	[1]450	14	.00047	Meandering	do	0.14 km upstream from gaging station.

[1]Estimated.

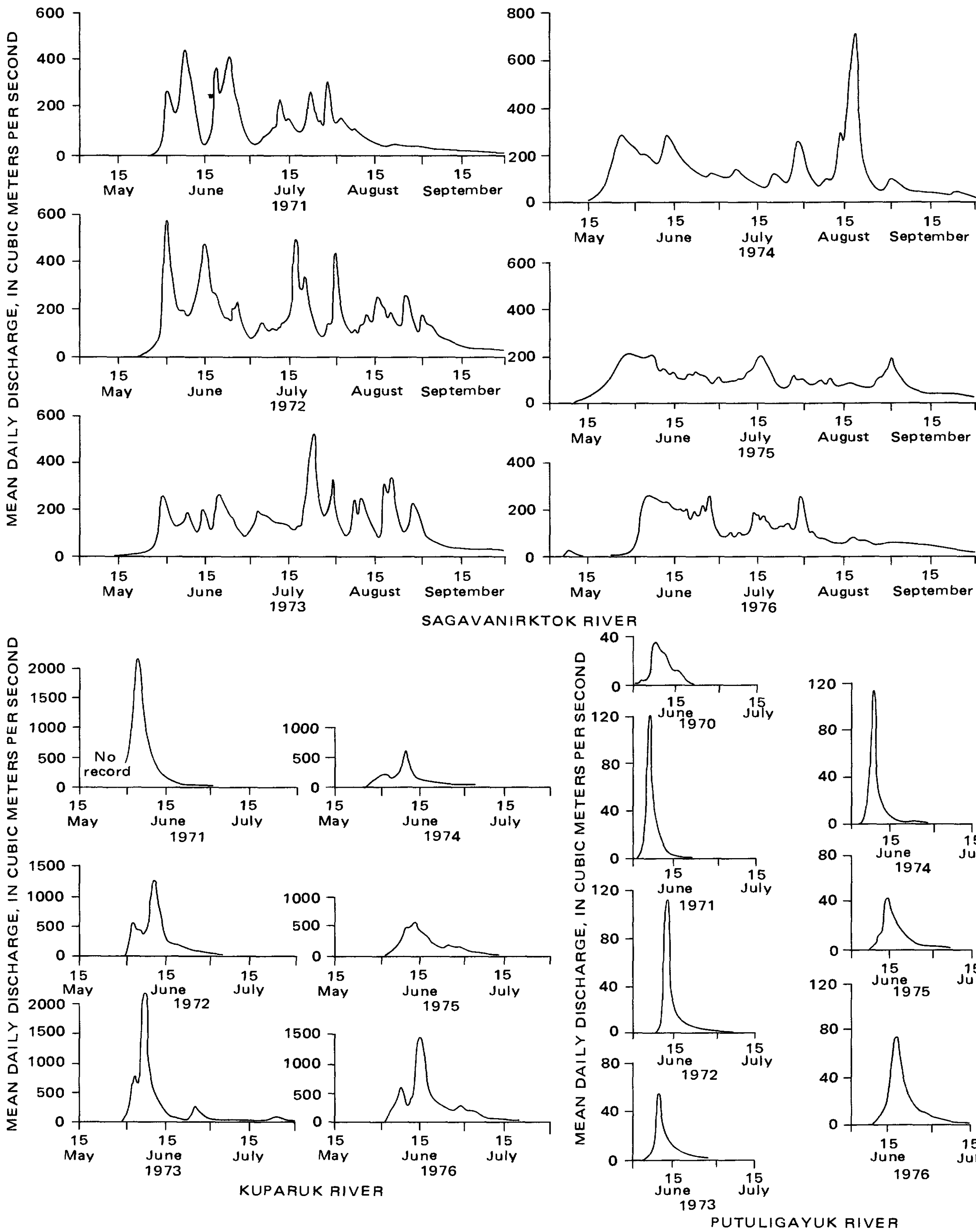

FIGURE 2.—Seasonal hydrographs of streams in arctic Alaska. Drainage areas at gaging stations: Sagavanirktok River, 5, 719 km²; Kuparuk River, 8,107 km²;Putuligayuk River, 456 km²; Discharge volumes include both water and sediment.

ATIGUN RIVER

A major headwater tributary of the Sagavanirktok River, the Atigun River occupies a broad glaciated valley in the Brooks Range. The upper section of the valley is a glacial outwash plain; downstream, the river meanders between generally cohesive banks in a valley flat where the study sites are located. The flat is the site of glacial-outwash deposition resulting from lake impoundment or reduced stream gradient caused by the late Pleistocene recessional moraines located north of Galbraith Lake. Channel slope in these reaches is now controlled by the head of a bedrock gorge into which flow of the Atigun River was diverted eastward to the Sagavanirktok drainage when its pre-glacial course as a tributary of the Itkillik River or Kuparuk River was disrupted by moraines.

Breakup of the Atigun River occurred in 1976 approximately on April 30. This unusually early breakup in comparison with that of nearby streams was in part due to the effect of wind-blown dust from the adjacent pipeline-construction corridor on the albedo of the valley snowpack. An initial peak on May 6 was followed by relatively constant lower discharges throughout May and then by successively higher peaks on June 5 and 12.

A major summer rainstorm occurred in the headwaters of the Atigun River valley on June 30, 1976. Measured storm precipitation was small (<40 mm) at a station within the valley (Atigun Camp) and at a station across the crest of the Brooks Range (Chandalar Camp). Nevertheless, local intensities must have been substantially greater in order to cause the significant runoff that resulted. The peak discharge exceeded that corresponding to bankfull flow throughout the Atigun River valley and, over 80 km downstream, caused the peak of July 30–31 on the Sagavanirktok River hydrograph (fig. 2).

SAGAVANIRKTOK RIVER

The major stream east of the Colville River in arctic Alaska, this river drains a large part of the Brooks Range. In the foothills the channel pattern is generally sinuous or meandering and the banks are cohesive, but downstream from the junction with the Lupine River, and beyond the maximum extent of glaciation, the channel becomes progressively more complexly braided and the banks more noncohesive. At site S4 the river splits into two branches, with the main channel entering the Beaufort Sea about 12 km east of the west channel.

A discharge of less than 0.1 m^3/s is characteristic of the Sagavanirktok River at the gaging station (fig. 1) through the winter months. Substantial flow over ice began before May 3 in 1976, reflecting the early breakup on the tributary Atigun River. The initial peak of over-ice flow on May 6, evident in the hydrograph of figure 2, was largely the result of this flow. Breakup in the channel between sites S1 and S3 did not occur until May 31, however, and breakup flooding did not peak in the same reaches until June 5. As has been historically typical of the stream since measurement began, flow remained at reduced but substantial levels throughout most of June and July in 1976, and rose to similar levels both on June 28 and in response to the summer rainstorm of July 30 in the drainage of the tributary Atigun River.

In two of six years the peak mean daily discharge on the Sagavanirktok River has occurred in response to rainstorms in July or August. In other years the more typical "nival" pattern prevailed, in which the breakup flood was the most significant flow of the year. The hydrographs (fig. 2) show that important summer storms have been relatively common in the watershed upstream from the gaging station.

KUPARUK RIVER

Although the headwaters of this stream are confined to the foothills of the Brooks Range, it is the largest drainage reaching the arctic coast between the Colville River on the west and the Sagavanirktok River on the east. The study sites are in the foothills where the Kuparuk River meanders across a 0.7-km-wide valley flat with cohesive alluvial fill and scattered areas of coarse glacial detritus in which sorted polygons have developed.

Flow over ice at the study sites began in 1976 approximately on May 2, but breakup did not occur until May 23. Breakup flow first peaked on May 26 and was followed by a second peak of nearly identical stage on June 6. The discharge was greatly reduced by the time of final observations on June 13. Much of this early flow in the upper foothills was not recorded at the gaging station near the mouth (fig. 2), probably because of storage in the form of icings at intermediate locations. A sharp temperature gradient existed northward across the coastal plain throughout May.

At the gaging station near the mouth of the Kuparuk River, the characteristic flow pattern since 1971 has been one of peaks in breakup flow during early June, followed by reduction in flow to low summer levels by late June. In 1976 breakup at the gaging station occurred on June 3 or 4, and the peak discharge was measured on June 15.

In contrast with the Sagavanirktok River, the Kuparuk River drainage yields much higher rates of runoff per unit area during breakup, and significant flow is largely concentrated in the June breakup period (fig. 2). Breakup occurs more uniformly in the Kuparuk River basin than in the Sagavanirktok River watershed because of the less diverse relief in the former. In addition, the runoff from summer rainstorms, which are most intense in the mountainous southern sections of the Sagavanirktok River basin, is greatly retarded by the tundra vegetation of the foothills and coastal plain once the active layer begins to thaw. The thick sod mat and hummock surfaces of the foothill and lowland tundra create both a great capacity for absorption of precipitation and a high resistance to flow. Thus, rainstorms in the Kuparuk River basin have not caused significant peaks during the period of flow measurement.

HAPPY CREEK

This small drainage of 89 km^2 enters the Sagavanirktok River from the west at Happy Valley Camp. It is typical of the foothill area; little valley flat is evident and the channel is broadly sinuous or meandering, with partial bedrock control. The bed material is coarse gravel, and the banks consist of a tundra mat over a thin veneer of gravel-rich soil.

Meltwater discharges in Happy Creek rise and recede rapidly. In 1976 breakup flow increased from 0 to 39.4 m^3/s between June 3 and June 6 and receded to low levels by June 8. Annual peak discharges before 1976 ranged from 2.5 to 16.2 m^3/s since crest-stage measurements began in 1972 (table 2).

PUTULIGAYUK RIVER

The drainage of this stream is confined to the Arctic Coastal Plain, an area of low relief with numerous oriented lakes similar to those so intensively studied in the vicinity of Barrow. The meandering channel has bed material of fine gravel and banks of cohesive silt and clay with soil development overlying fine gravel.

Measurements since 1970 indicate that the Putuligayuk River generally peaks rapidly, rising from zero flow during 1 to 2 weeks in early June and falling continuously to low summer levels in about the same length of time. The same highly attenuated summer flow regime that is characteristic of the Kuparuk River basin is present to an even greater degree in the Putuligayuk River basin. To the hydrologic effects of the tundra are added the effects of poorer drainage and the presence of standing water in a greater proportion of the watershed.

Summer storms historically have caused little runoff; through 1976 no discharge in excess of 3 m^3/s has been recorded following recession of the breakup flood. Significant flow, mainly over ice, began in 1976 approximately on June 5, and the peak discharge occurred on June 18. Observations of the channel were ended on June 12.

TABLE 2.—*Peak discharges measured at crest-stage gage on Happy Creek. Drainage area above gage is 89 km^2*

Year	Date	Discharge (m^3/s)
1972	Unknown	14.9
1973	Unknown	2.5
1974	During May	[1]11.3
1975	June 6	16.2
1976	June 6	39.4

[1]Estimated; ice effects.

DATA COLLECTION

MEASUREMENT TECHNIQUES

Erosion rates in cohesive bank materials were measured with erosion pins (6 x 178 mm) emplaced horizontally at appropriate intervals in a bank cross section (see Wolman, 1959, p. 207). The sites selected were those at which maximum amounts of bank erosion could be expected, such as the outside of a meander bend and other locations where nearshore current velocities were high and banks were known from field evidence to be eroding. Erosion rates at some sites were sufficiently high to require the periodic replacement of a few pins.

Erosion rates in noncohesive bed and bank materials were determined in a similar manner through measurements of the channel surface relative to lengths of reinforcing bar (9.5 mm x 1 m) driven to depths that penetrated frozen material. Scour measurements were made in the same way. Horizontal control was established by tape survey at the time of emplacement; vertical control was made by referencing the tops of the reinforcing bars and the bed surface to stage, which in turn was recorded with improvised staff gages. As the noncohesive beds and banks thawed during the study period, it was necessary periodically to drive the bars deeper into frozen material, each time referencing the new height for vertical control.

Rates of migration of the frozen-unfrozen interface were obtained by driving lengths of reinforcing bar into the bank or bed at the location of each erosion pin or bar. This procedure worked well after verification by excavation and repeated measurements. With practice it was possible to distinguish the nature of whatever rigid surface was encountered—whether frozen bed material, segregated ice, or a large clast—by the "feel" of the probe when driven against the obstacle. The difficulties described by Mackay (1977, p. 327) in probing the depth of the active layer in fine-grained soils with little ground ice were not encountered. In both cohesive and noncohesive sediment the boundary could be determined accurately. Where the bed material was too coarse to use a probe, thaw rates were determined by successive excavations that were offset sufficiently to avoid any thermal disturbance from the preceding excavation.

RATES OF THAW IN BEDS AND BANKS

The variation in thaw rates with sediment size was determined in a section of the Atigun River along which bed material changed in size but discharge and thermal conditions were relatively constant. Figure 3 shows the variation in thawing rate of submersed bed and bank material plotted with change in midday water temperature. Each data set represents the conditions at a single erosion pin or bar in the channel, and one that is typical of many similar sets of data, each corresponding to the location of a pin or bar in the cross section. Sediment in the silt (0.004–0.0625 mm) and clay (less than 0.004 mm) size ranges at site A2 is bank material; the gravel (over 2 mm) at site A1 and the sand (0.0625–2 mm) at site A3 are bed material. At the latter two sites major sections of the banks are composed of the same material as the beds, and the submersed thaw rates in those parts of the banks were similar to the rates measured in the beds.

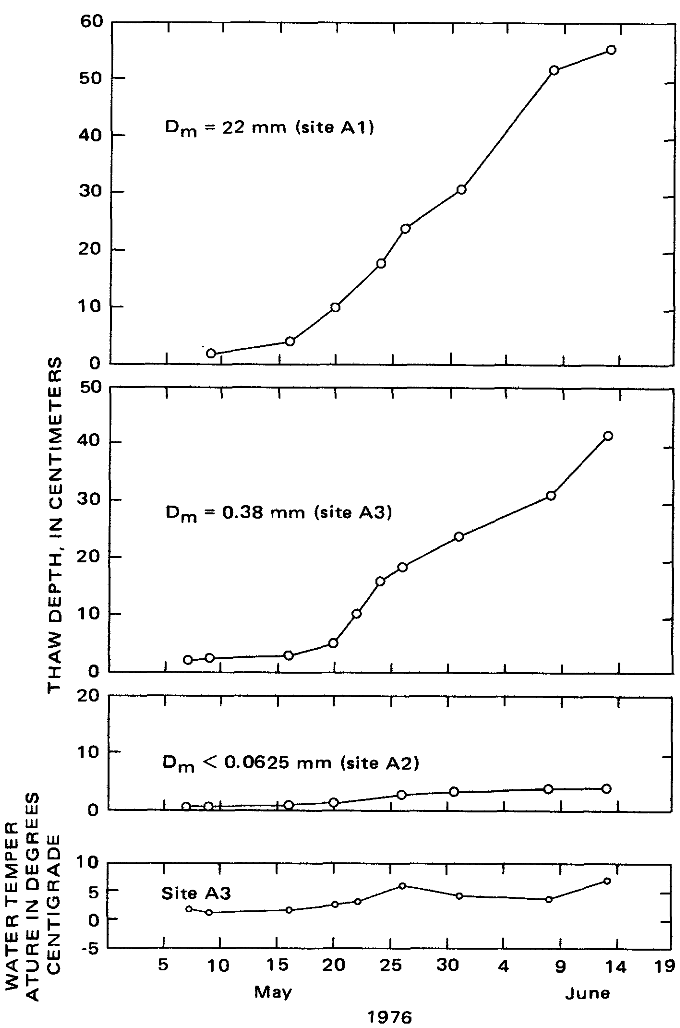

FIGURE 3.—Thaw depths in submersed bed and bank material of differing texture at three sites on the Atigun River. Water temperature was measured at site A3. Size was determined by field measurements at site A1, by sieving at site A2.

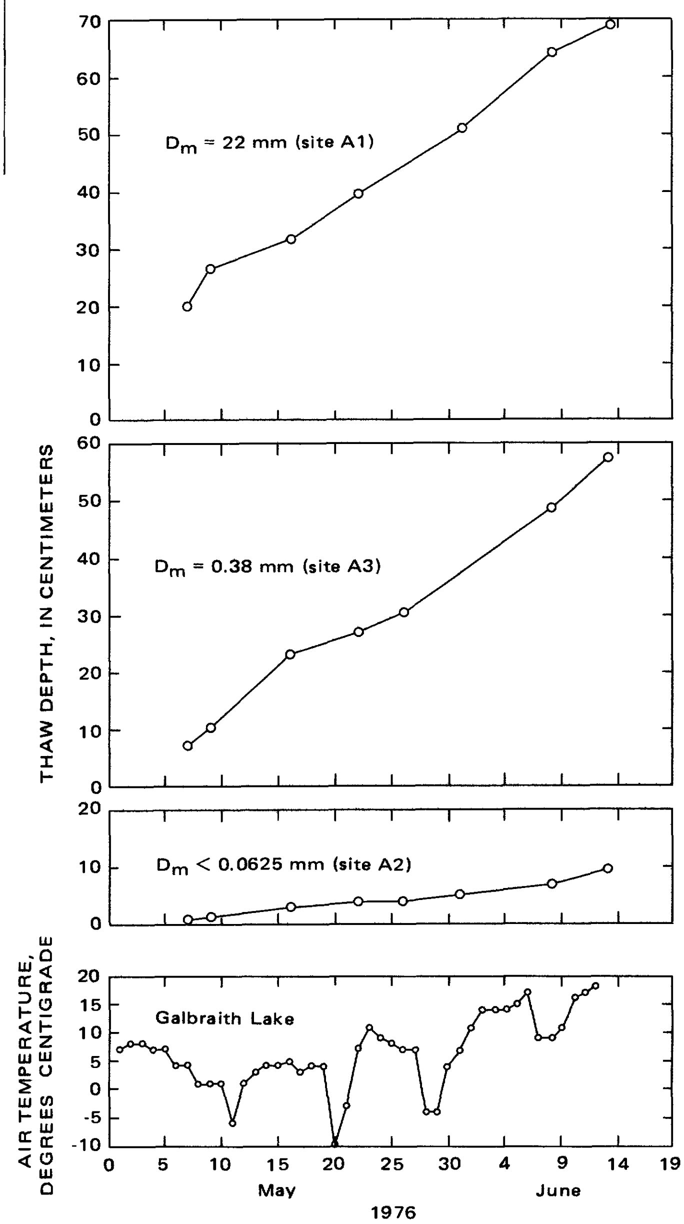

FIGURE 4.—Thaw depths in emersed bed and bank material of differing texture at three sites on the Atigun River. Air temperature was measured at Galbraith Lake.

A similar set of measurements was obtained for emersed sediment of the same sizes at the same three locations. Each point (fig. 4) is also typical of a group of points at each site. Air temperatures shown are the daily maxima at Galbraith Lake, 10 km northwest of site A3.

Texture proved to be the dominant factor in the variations in thawing, given the approximately equal flow and thermal conditions at the three sites. Thaw rates were proportional to grain size whether sediment was exposed to water or air. In the absence of other factors, the effects of the latent heat of ice should produce the lowest thaw rates in sediment with the greatest ice content. That this was not seen probably reflects the efficiency of the finer, more cohesive sediment, once it has thawed, in insulating the frozen interface and thereby retarding further thawing.

Position in the cross section—differing flow depths and whether the measurement was in bed or bank material—was of relatively minor importance in explaining variations in thaw rates of submersed sediment. Emersed thaw rates proved to be moderately greater than submersed rates for each of the three textures observed during the period. When occurring coincident with erosion, however, the potential for much greater rates of thawing in submersed material is obvious.

Although air temperature trended downward between May 1 and 21, the thaw depths in emersed, noncohesive material increased rapidly during that period. It is likely that incident solar energy increased throughout that time interval. At only one point at one site (A3) was any decrease in thaw depth observed—in this case associated with the unusual cold spell of May 20–21. At other points the active layer thickened continuously.

THAW AND EROSION RATES COMPARED

The question of permafrost control of stream behavior hinges on the relative amounts of thaw and erosion at individual points. If erosion uniformly proceeds more slowly than the observed rates of thaw, for example, there is clearly no direct effect of frozen bed or bank material on erosion, and thus on stream behavior. If frozen bed or bank material is in contact with the flow after breakup, it can be assumed that erosion is being retarded by the frozen condition of the material.

Figures 5 and 6 summarize the measurements of thaw and erosion, with the data divided according to whether the material is cohesive or noncohesive. The distinction in sediment type is based on field examination. The measurements are confined to the time interval between breakup and the peak of breakup flooding in each stream. A large number of points representing no erosion and minuscule amounts of thaw are not shown. Such points are typical of the smaller streams.

The results show that the general circumstance is for erosion to proceed more slowly than thaw, with the exception of two specific locations where sediment was

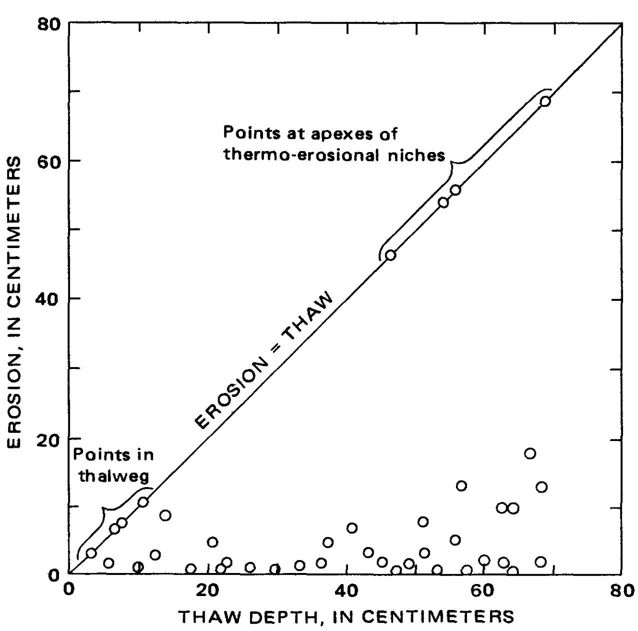

FIGURE 5.—Plot of erosion versus thaw for points in noncohesive bed and bank material. Points without detectable erosion (41) are excluded.

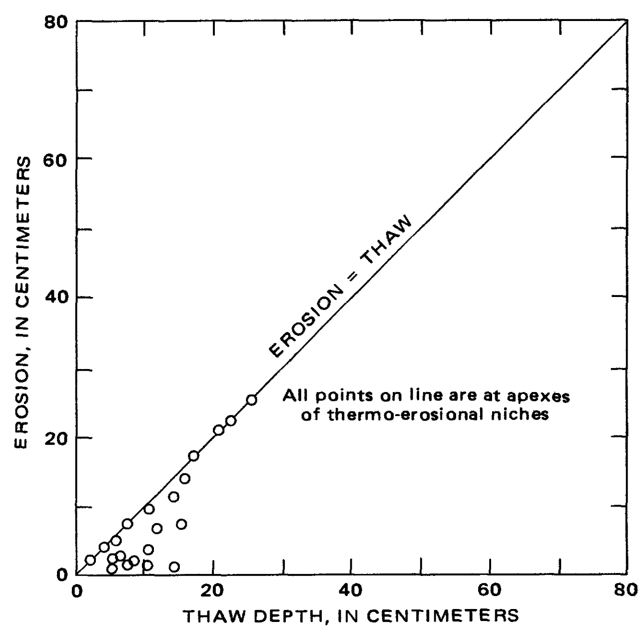

FIGURE 6.—Plot of erosion versus thaw for points in cohesive bed and bank material. Points without detectable erosion (25) are excluded.

frozen at the bed surfaces. These were (1) the thalwegs of streams—sites of the largest amounts of vertical scour; and (2) the innermost surfaces, or apexes, of thermo-erosional niches—sites of the most intense lateral erosion. Most such points were not frozen; yet the only points that remained frozen after breakup on the larger streams were at those locations. In plotting figures 5 and 6, it was assumed that the retreat of a solidly frozen surface represented equal amounts of thaw and erosion.

That maximum rates of erosion are less in cohesive material than in noncohesive material is evident from a comparison of figures 5 and 6. This difference is not the sole function of the lesser rates of thaw in cohesive sediment seen in figures 3 and 4. Were this the case, control of erosion by permafrost in cohesive banks would be greater than in noncohesive banks, and in figures 5 and 6 the points representing cohesive sediment would be located nearer the line representing equal amounts of thaw and erosion. Although a comparison of the figures suggests that permafrost control of erosion is more likely in cohesive sediment, the greater resistance of the finer sediment to erosion must also be attributed to the properties that cause that size of bed and bank material to thaw more slowly—the low permeability that produces low rates of thawing is a function of the cohesive bonding of clay minerals that increases the resistance to erosion.

Measurements on the larger streams (Sagavanirktok River sites S1–3 and all Atigun River sites) showed that direct permafrost control of erosion in most cases ended well before the peak discharge occurred. The situation was much different for smaller streams like the upstream Kuparuk River and Happy Creek, where frozen material remained at or near the surface, especially in cohesive bank material, after recession of the breakup flood.

THERMO-EROSIONAL NICHING

Bank undercutting that can be characterized as thermo-erosional niching is the dominant cause of bank retreat in the streams with banks of mainly cohesive material, a category that includes nearly all streams with sinuous or meandering channels in the study area. In typical thermo-erosional niches in homogeneous deposits, the top of the niche develops at sustained levels of high flow. The development of niches is discouraged by heterogeneity of banks and by fluctuations in stage. Although the bank undercutting in most study streams is less regular than that illustrated from a noncohesive bank (fig. 7*B*), both vertically and longitudinally, the process of bank undercutting by concomitant thaw and erosion is the same. It should be noted that bank undercutting and subsequent failure are unique in the arctic setting only in their prevalence and degree of development. The increased strength provided by permafrost in the banks permits greater amounts of undercutting, which in turn produces larger slump blocks.

The niching process is more rapid in the layers of sand and gravel that are sporadically interbedded in cohesive banks (fig. 7*A*), as would be expected from the greater rates of thaw and erosion in the coarser sediment. Where banks are uniformly of silt and clay, niche development may be concentrated at the base of the cohesive fill, within the underlying gravel. A corollary to these observations is that the relative susceptibility of cohesive banks to erosion can be assessed from the stratigraphy of the alluvial fill, observable either in bank cross sections at times of low flow or by excavation.

Niching is also present in braided channels. Where channels braid in predominantly sand-size material, the niches are developed most dramatically (fig. 7*B*) and correspond in morphology to the previously described examples (Gusev, 1952; Walker and Arnborg, 1966). In small channels in sandy reaches of the Atigun River an entire anabranch was seen to disappear within a niche at some sites. With coarsening of bed and bank material above the size range of sand, niching becomes progressively less well developed. Consequently, the process is of lesser importance in the braided streams than in those with sinuous or meandering patterns.

BANK SLUMPING AND VEGETATION EFFECTS

The failure of cohesive banks in response to undercutting is generally a gradual process over hours or

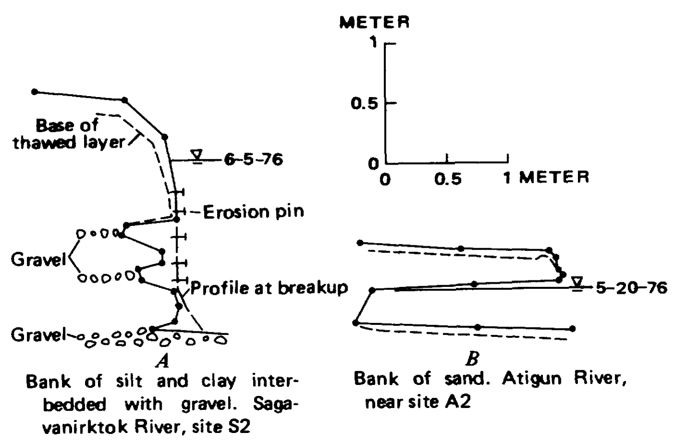

FIGURE 7.—Cross sections of undercut banks. *A*, Cohesive. *B*, Noncohesive. Dates indicate water levels at time of survey.

several days. Sloughing of small blocks from the vertical face of noncohesive banks (Outhet, 1974, p. 307) is an important mode of failure as is the sudden collapse of both types of banks upon fracture through frozen material weakened by thawing (Walker and McCloy, 1969, p. 76). True slumping is not particularly common, but the term may be appropriate, as a simplification, for the assemblage of mass-wasting processes involving bank failure. Failures probably tend to be accelerated during flow recession (Inglis, 1949, p. 152), by the reversal of seepage forces and the loss of the resisting pressure provided by flow in the channel.

Ice wedges associated with the raised-edge polygonal ground on the coastal plain may thaw rapidly and provide planes of weakness along which bank failure can occur. Segregated ice masses thaw more rapidly than most cohesive sediment but appear to thaw less rapidly than some highly erodible noncohesive sediment. This seeming contradiction to the effects of the latent heat of ice on thaw rates probably reflects the ability of the cohesive sediment to insulate the thaw interface. Ice wedges occur in all flood-plain deposits of the study

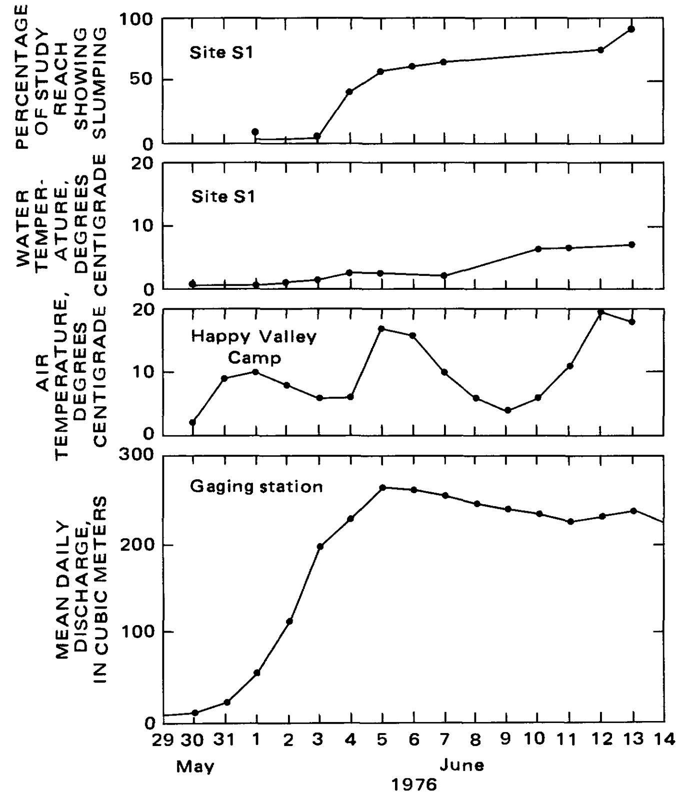

FIGURE 8.—Plot of percentage of bank failures at site S1, water temperature at site S1, maximum daily air temperature at Happy Valley Camp, and mean daily discharge at gaging station on the Sagavanirktok River. Discharge records are estimated for the period shown.

area but are of quantitative importance only in the coastal plain.

The bonding provided by the tundra root mat can significantly retard slumping. It is also possible, because cold climates retard the decay of roots, that lateral erosion may be retarded directly by root structures to an unusual extent (Smith, 1976). Because the shallow root depth of the modern tundra vegetation is limited by the depth of active layer (less than 0.3 m in vegetated cohesive material), roots are not important in limiting bank erosion in the larger streams of the study area. Older organic layers at depths up to 1.0 m are preserved in a frozen condition and may retard erosion directly in the lower banks of the smaller streams. Bank erosion was limited by vegetation during the record discharge in Happy Creek, but other important factors were sediment size and depth of thaw. Vegetation on the banks of Happy Creek was important in the indirect sense that it greatly increased the insulation properties of the bank surfaces and thereby decreased both the thaw rate and subsequent erosion.

The direct role of vegetation in most study streams was mainly that of retarding the failure of undercut banks to varying degrees. The tundra mat was commonly observed to remain attached to the bank, folded downward into the stream on the top of a failed bank and thus acting temporarily to prevent further erosion, as described by Walker and McCloy (1969, p. 74).

VARIATIONS IN TIMING OF BANK SLUMPING

The percentage of the reach showing slumping of the bank at site S1 on the Sagavanirktok River is plotted with time in figure 8, along with discharge and associated air and water temperatures. The sudden increase in slumping seen on June 4 was related to niching that was accelerated by rising water temperatures on June 3 and 4. There was clearly little time lag between the niching and bank failure at this site.

The importance of undercutting and subsequent bank failure in the study streams varies with size of the drainage area when the comparison is limited to a single channel pattern and sediment type—that is, to sinuous and meandering channels with banks of cohesive sediment. Figure 9 shows the prevalence of slumping along particular sinuous and meandering reaches within the time period between breakup and the breakup peak. Each reach represents approximately the outside bank of a single meander bend between the points of inflection.

Figure 9 indicates that the rise of breakup flooding was vastly more important on the Sagavanirktok River than on even the Atigun River, the next largest. No bank failures of any kind were observed in association with breakup flooding on Happy Creek, the smallest drainage studied. And, at the Kuparuk River sites only a few small failures were initiated near the end of the period. Limited observations on the Putuligayuk River revealed only a slight amount of bank slumping beginning as of June 12, and similarity in pattern and bank material suggested the likelihood of behavior like that of the study sites on the Kuparuk River. The causes and implications of this variation with drainage area are discussed at length in a subsequent section.

Referencing the data of figure 9 to the period between breakup and the peak of breakup flooding provided a geomorphologically and hydrologically significant time interval over which to compare the importance of slumping in different watersheds. The amount of slumping proved to be a valid index of lateral erosion. Had absolute values of lateral erosion been averaged in each sample reach and plotted in figure 9, the results would have been similar, even though the maximum amount of erosion and the proportion of reach affected by that erosion do not necessarily correlate. The length of reach affected by slumping is a variable easily determined with successive ground photographs, compared to values of total erosion that can be determined only at surveyed cross sections or with aerial photography.

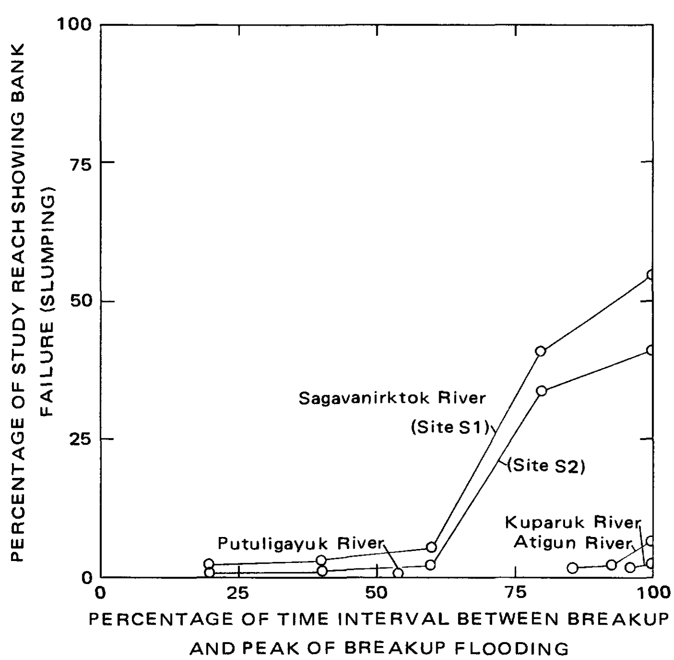

FIGURE 9.—Plot of percentage of sample sinuous or meandering reaches showing bank failure, versus percentage of time interval between breakup and peak of breakup flooding. Only partial data were collected from the Putuligayuk River.

In appraising figure 9, it is worth noting that the time lag between undercutting and slumping of a bank can vary with a number of factors. It was verified, however, that the lack of slumping at the sites in the smaller watersheds was due to an actual lack of lateral erosion and not to a time lag between niching and failure.

SAGAVANIRKTOK RIVER, SITES S4 AND S5

It was noted in the introduction that cohesiveness, as a function of the particle-size distribution of bed and bank materials, is the main factor in determining bank erodibility, and that erodibility in turn can be considered to be the major variable in determining channel pattern. Consequently, the association between channel pattern and erosion rates corresponds to that between bed material and erosion rates—the greater erosion rates occur in braided channels, which are characteristically formed in noncohesive sediment. Because of anomalous local conditions, downstream reaches of the Sagavanirktok River were exceptions to both this relation and the variation with drainage area.

The time of breakup in watersheds north of the Brooks Range is the reverse of that expected from their altitude, with the lowest and northernmost basins being the last to yield flow. Dense coastal fog remained longer than usual in 1976 and kept temperatures low through mid-June in an area as far as 60 km from the Arctic Ocean. As a result, the response of channels to breakup flooding in the major, through-flowing rivers depended, as it would to a lesser degree in more typical years, on local conditions as well as on the thermal input of flow from the earlier breakup farther south.

From June 5 to 12 the bank material and most bed material in braided reaches on the lower Sagavanirktok River (sites S4 and S5) remained frozen although the peak of breakup flow had passed on June 5 or 6. On June 11 at site S4 there was little evidence of any channel erosion, and neither submersed nor emersed sediment had thawed more than several centimeters in depth. Flow depth at site S5 prevented measurements in the deeper channels, but in observable respects the situation was similar to that at site S4. The high upstream water temperatures (5° to 7°C at sites S1–S3) were rapidly dissipated in the shallow, complexly braided channels at these two sites on the fog-bound coastal plain. Water temperature was only 0.6°C at site S4 at the same time it was over 6.0°C upstream at site S1 (fig. 8).

This variation in channel response with climate illustrates the type of local complication that can dominate any general relations like those based on bed material or drainage area. It also emphasizes the major variations that are possible along the larger drainages that cover considerable latitude in the Arctic.

BED SCOUR

The effect of frozen bed material on scour was studied with repeated profiles of the bed and thaw surface across the channel at site A3 on the Atigun River. Bed material at this site is fine sand (median diameter=0.38 mm) within the size range that shows large increases in bed-material discharge with low water temperatures (Colby and Scott, 1965, fig. 19). The channel at that point had a sinuous pattern during the observation period, with tendencies toward braiding at the lower end of the range of observed discharges. Throughout most of the observation period, a single channel from 25 to 30 m in width and with a depth of less than 1 m was present.

Figure 10 portrays the channel change and the increases in thaw depth between May 20 and 31. Before May 26 little lateral movement occurred, and the bed cross section was about the same on May 7 as that shown on May 20 and 26. As late as May 20, frozen material was sporadically exposed at the bed surface in the thalweg (point a in figure 10). Sediment transport during the pre-May 20 period was evident in the form of dunes moving across a frozen substrate.

The importance of figure 10 is that scour was being retarded by frozen bed material as late as 3 weeks after breakup when midday water temperatures had risen as high as 2.5°C (fig. 3). That the scour did not increase once the thalweg had thawed to greater depths, shown by the profiles of May 31, may reflect the greatly increased load in the stream that accompanied the thawing process and was promoted by the combination of fine-grained bed material and low water temperatures.

Figure 10 also shows the sudden increase in lateral mobility of the channel that occurred after May 26 and the corresponding change in configuration of the thaw surface. These changes occurred under conditions of relatively constant stage, excluding small diurnal changes.

CHANNEL PROCESSES DURING REMAINDER OF 1976 RUNOFF SEASON

The study sites were resurveyed in May 1977 to determine the channel changes that occurred during that part of the 1976 runoff season following breakup flooding. At the time of the 1977 survey, breakup either had not occurred or was in an early stage, and the channels were as they existed at the time of freezeup in September 1976. The resurvey was surprising in the

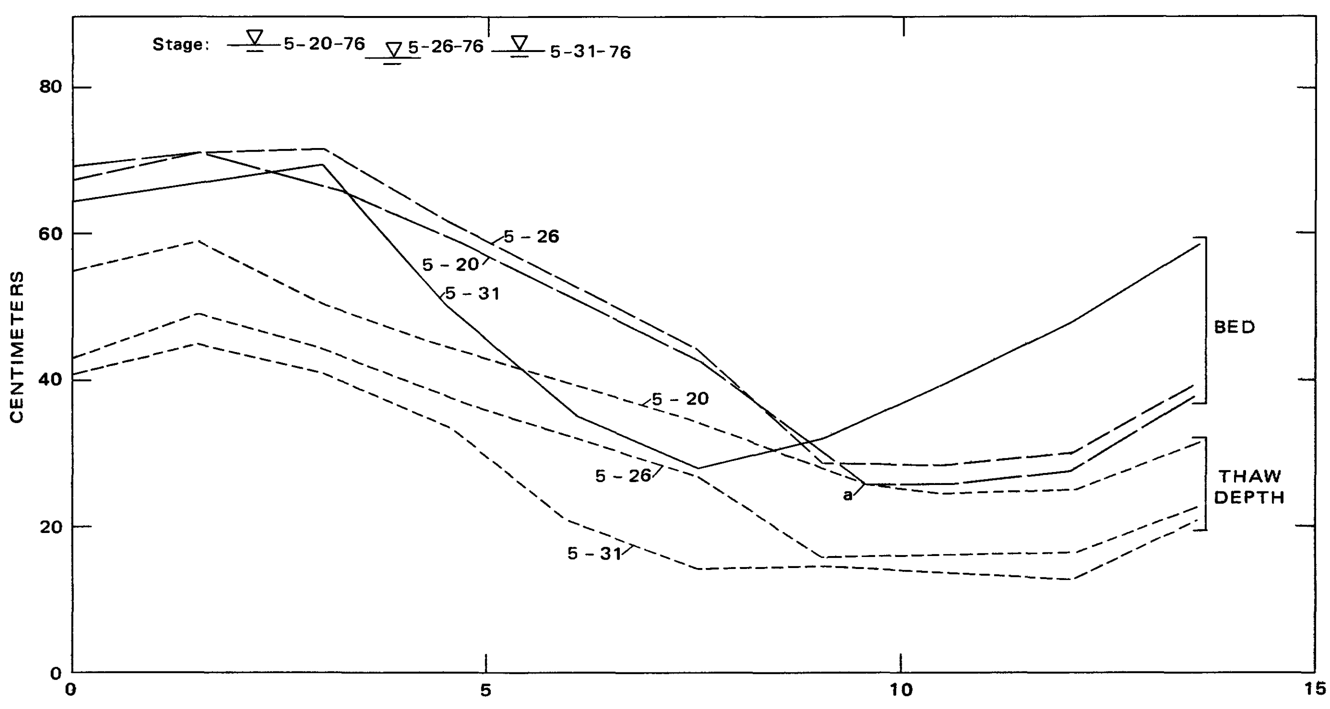

FIGURE 10.—Cross sections of bed surfaces and thaw surfaces in the thalweg at Atigun River site A3. Stages indicated are water levels on dates of survey.

small-to-nonexistent degree of change that was observed to have occurred at most sites in 1976 after at least partial recession of the breakup flow.

The absence of change was striking in the smaller watersheds like Happy Creek and the upstream Kuparuk River. In the latter watershed individual slump blocks had not moved detectably from positions recorded by photography following recession of breakup flooding in June 1976. Slumping observed along the Putuligayuk River was comparable to amounts recorded during the breakup period on the Kuparuk River (fig. 9), but, because observations were not continued through the complete period of breakup flooding on the Putuligayuk River, the failures there could not be ascribed to the flooding period or the remainder of the runoff period.

The exceptions to the observations of little change were the sites on the Atigun River and the downstream part of the Sagavanirktok River.

ATIGUN RIVER

The rainstorm flood of July 30 was a major instrument of lateral and overbank erosion throughout the entire length of the Atigun River valley. The peak flow of the July 30 flood exceeded bankfull discharge throughout the observed sections of the river. It locally approached, but did not exceed, the levels of the maximum evident flood measured by Childers and Jones (1975, table 1 and p. 16–17). On that basis, it can be estimated that the peak discharge was about 300 ($\pm$50) m^3/s in the reaches near site A1. The recurrence interval of the flow could be approximated within the range of 5 to 15 years.

Amounts of lateral erosion were determined by comparing ground photographs at the study sites and in nearby reaches. Lateral erosion was as much as 4.9 m in the case of a terrace level 1.4 m above the flood plain. The banks were eroded by a measurable amount at most locations where erosion would be expected—outsides of bends and, in straighter reaches, points of impingement of the thalweg against channel banks. Overbank flow on vegetated areas of the flood plain locally stripped the tundra mat from surfaces shallowly underlain by gravel.

An impressive feature of the flood results was the resistance to erosion of uniform cohesive deposits, especially where a cutbank was absent and vegetation extended to the level of the low-water channel.

SAGAVANIRKTOK RIVER

The situation at sites on the sinuous or meandering portion of the channel (S1 and S2) in May 1977 was similar to that observed after partial recession of breakup flooding in mid-June 1976. Some movement of individual slump blocks had occurred. Only a few had been removed by erosion, and many remained folded down into the channel with the intact tundra mat acting to prevent subsequent erosion. The small amount of change was unexpected, particularly so in light of the fact that flow continued at significant levels throughout much of the remaining runoff season (fig. 2).

Site S1 was observed daily for a week during the rise of the breakup flood in 1977 (May 25–June 1). During that time it was apparent that the previous year's slump blocks were acting to inhibit lateral erosion. This result raises the possibility that more than one year may be necessary to remove the slumped sections of a bank and prepare it for another large increment of lateral erosion by means of niching and failure.

The preceding sequence of large change during breakup flow and little change during the rest of the runoff season was reversed at the sites on the braided sections of the river downstream (S4 and S5). There, measurable change in the position of most of the individual anabranches had occurred, indicating that the portion of the runoff season after recession of the breakup flood is probably the normal time of channel change. Channels at these sites were mainly frozen and unchanged throughout the breakup flooding of 1976.

SEDIMENT TRANSPORT

With the frozen condition of the channels seen to last at least part of the way through breakup flooding, there are implications for the measurement and estimation of sediment discharge in arctic streams that involve questions of sediment availability. Sediment in transport has been divided conceptually for practical purposes into two components: the wash load (or fine-material load)—consisting of fine sediment the discharge of which reflects events upstream and is considered not to be a function of flow hydraulics; and bed-material load (or bed-sediment load)—generally coarser sediment of sizes present in the bed, the discharge of which is considered a function of flow hydraulics. The wash load in this concept must be measured by sampling; the bed-material discharge is commonly estimated for engineering purposes with relations involving flow and sediment parameters. If the bed is partly frozen, however, sediment may not be readily available at a point and the total load may thus become, like the wash load, a function of conditions and sediment availability upstream. This means that for some arctic streams at certain times the total sediment load can be determined only with measurements.

The effect of a frozen bed on sediment transport will be greatest for the coarsest fractions of bed material—those sizes most likely to move only as bedload, rolling or skipping along the bed surface. It will certainly be most noticeable in the smaller streams in which the frozen condition of the bed may extend through the entire breakup period. If such an effect is significant in the larger streams outside the coastal plain, it will probably be confined to the initial rise of the breakup flood. Although the coarse nature of the sediment (table 1) made direct measurement of bedload impractical, several lines of evidence can be used to determine first, when bed material was thawed enough to be mobile during the 1976 breakup flooding, and second, whether breakup flows were competent to move the available material.

At sites S4 and S5, on the Sagavanirktok River, the effects of a predominantly frozen bed and bank probably had considerable influence on sediment transport through the rise and for some time after the 1976 breakup flooding. Thaw depths measured at upstream sites (S1–3), however, indicated that the bed was probably capable of becoming mobile within 1 or 2 days after breakup. Consequently, bed-material transport there was mainly a function of the competency of the flow, and was entirely so later in the rise of the breakup flood as competence increased to the point where transport of some sizes present in the bed was possible.

On the Kuparuk River, observations indicated that no movement of cobble-size (64–256 mm) bed material accompanied the first breakup peak of May 26. Soundings with a probe during and the day after the peak revealed directly that cobble-size clasts were frozen in an immobile bed. However, extensive transport of the same material occurred during the second breakup peak of nearly equal discharge, thereby suggesting that the competency of the first flow was sufficient to cause transport. Both peaks were approximately equivalent to bankfull discharge.

The sand-size bed material of the Atigun River at site A3 was continuously transported to some degree throughout the breakup period. Data were inconclusive on whether transport of the gravel at site A1 occurred before the peak of June 5, but the depth of thaw would have permitted movement had flow been sufficiently competent. Extensive movement of the coarser material occurred with the peaks of June 5 and 12.

A small amount of bed-material transport occurred with the sudden rise and sharp peak of breakup flooding in Happy Creek. The peak discharge of 39.4 m^3/s was the maximum recorded since crest-stage measurements began in 1972. Transport of gravel near the lower end of the pebble-size range (4–64 mm) could be observed in culverts below Happy Valley Camp. Coarser sediment was present in the bed yet was not transported in spite of the fact that flow near the peak was greatly in excess of that competent to do so, based on estimates of flow depth and slope and the recognized relations between size and shear stress on the bed (Baker and Ritter, 1975, fig. 1). Although observations of the bed during the short interval of the peak were not possible, it seems reasonable to suggest that the coarser material was frozen in place and that thaw in response to flow released only the finer gravel. Such appeared to be the case after flow receded.

DATA INTERPRETATION

EFFECTS OF PERMAFROST

Variations in thaw and erosion rates with size of bed material have obvious implications. Banks that are dominantly of cohesive material will erode more slowly than those with noncohesive sediment. More practically, the relative stability of streams with cohesive banks can be assessed by both the degree of interbedding with noncohesive sediment and the thickness of the cohesive fill relative to the depth of the channel. These measures reveal the susceptibility of a bank to thermo-erosional niching, the main mechanism by which the cohesive banks of artic stream channels retreat.

Permafrost was seen to retard bank erosion and scour in all the study streams; that is, frozen sediment was observed in direct contact with flow in channel thalwegs and at the apexes of thermo-erosional niches after breakup. The extent and duration of the effect, however, was strongly related to the size of the drainage basin when the observations were confined to sinuous and meandering channels with cohesive banks.

Data from all sites on the Kuparuk River and Happy Creek indicated that, if rise of the breakup flood is rapid and involves the lesser discharges and velocities of the smaller streams, little or no lateral erosion and scour may occur and bed-material discharge may effectively be prevented by the frozen bed. The data from the two upstream sites (S1 and S2) on the Sagavanirktok River show that, if the breakup flood involves the higher discharges and velocities of the larger streams, lateral erosion and probably scour as well may be extensive. Bed-material discharge is probably only briefly retarded after the start of breakup flooding in such streams. During the lengthy, six-week interval between breakup and the peak of flooding in the Atigun River, a stream of intermediate size, approximately the first half of the interval was characterized by frozen bed material and the second half by general channel mobility.

The variation in lateral erosion and thaw rates with watershed size reflects a basic variation of those processes as functions of stream power (proportional to the product of discharge and slope). The net result is doubtless highly time dependent as well, reflecting the shorter duration of the flooding in the smaller watersheds and the commonly lower water temperatures observed during breakup flooding in those watersheds. Watershed size is merely an easily determined factor that explains the variation in the sense of an independent variable in a regression analysis.

That there are definite differences in stream behavior related to size is indicated by the almost total absence of bed and bank processes before the breakup peak on the smallest streams. This observation is significant in the case of Happy Creek because the peak discharge in 1976 was by far the largest of record (table 2), suggesting the improbability of contradictory results in previous years. In streams like Happy Creek, the critical level of thermal input necessary for any significant bed or bank processes may not be reached before the breakup peak or even during the breakup flooding. It is likely, therefore, that the channels of the smaller streams respond significantly only to major floods, in contrast to those of the larger streams which plainly are modified by the annual breakup flooding.

A corollary to these conclusions is the effect of permafrost on lateral erosion once the peak of breakup flooding is past. In most of the larger channels thaw depths had increased to the point where little direct effect of permafrost on bank erosion or scour would occur in the absence of flood flows during the remainder of the runoff season. In fact, the effect of permafrost during the period after the breakup peak may be to facilitate the erosion of cohesive banks by assuring saturation of the thawed material. Moisture content was found by Wolman (1959, p. 211–212) to be the main factor not directly related to stage that increased the erosion of cohesive banks. Permafrost increases the moisture content of banks in two ways—the direct thawing of permafrost acts as a continuous source of moisture through the summer, and the base of the active layer functions as a barrier that will prevent the subsurface infiltration of runoff.

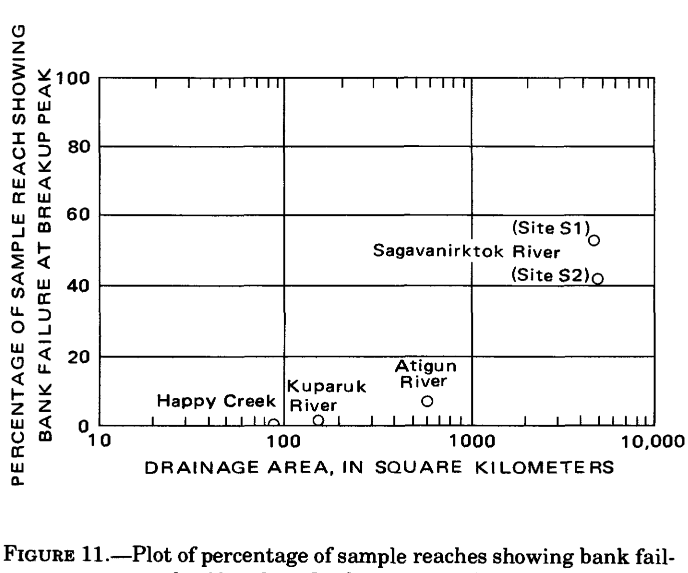

FIGURE 11.—Plot of percentage of sample reaches showing bank failure at peak of breakup flooding, versus drainage area.

THE VARYING IMPORTANCE OF BREAKUP PROCESSES

The plot of drainage area and the proportion of a study reach affected by slumping before the peak of breakup flow (fig. 11) illustrates at least a partial answer to why investigators have ascribed such different importance to the breakup period in terms of channel processes. That the most visible channel process in all arctic streams—bank slumping—can vary greatly with the size of the stream is a probable component of the explanation.

As noted in the introduction, a complete spectrum of conclusions as to what part of the runoff period is the most significant has been obtained. The results of study of large streams like the Colville River (Walker and Arnborg, 1966), the Mackenzie River (Outhet, 1974), and the upstream channel of the Sagavanirktok River in the study area would suggest that breakup processes are the dominant erosional events of the entire runoff period. At the other extreme, results from the relatively small drainages of Banks Island (Miles, 1976) or the smaller streams of the study area would indicate that breakup flooding was of comparatively minor importance.

Obviously, a simple correlation like that of figure 11 is not the complete answer to why some stream channels change significantly during breakup and others do not. Variation with other factors may obscure that relation. The effect of climate in the case of the sites on the lower Sagavanirktok River is an example of only one such possibility, in which channel changes on a large stream could be slight during breakup but pronounced thereafter. One can envision a similar set of circumstances that might even delay significant channel change until the later stages of the summer runoff period. Slumping on the lower Lena River is in fact concentrated in the late summer (Abramov, 1957), but this may only reflect a time lag between niching and slumping (Czudek and Demek, 1970, p. 105).

SUMMARY AND DISCUSSION

The preceding discussions provide first approximations of answers to the two problems posed at the beginning of this paper. The results are encouraging in that the seemingly contradictory results of previous investigation can be at least partly reconciled by the indicated variations in timing and intensity of arctic stream processes with bed-material size, channel pattern, drainage area, and climate. It is logical for investigators to have expected that permafrost would be a simplifying element in the study of arctic stream channels by possibly masking or dominating the effects of other variables. This expectation naturally led to contradictory generalizations on the basis of results from a single stream or from similar streams. The results from this study suggest that permafrost is best considered as no more than an additional variable, albeit an important one, in the study of arctic streams.

Three additional and closely related aspects of erosion in arctic streams are (1) the effects of major floods, (2) long-term erosion rates, and (3) the hypothesized uniformity of arctic erosion rates.

EFFECTS OF MAJOR FLOODS

We do not have sufficient flow records or erosion measurements from the Alaskan Arctic to assess the relative effects of major floods versus annual events with any certainty.

No truly outstanding flood has occurred in the streams with gaging stations during the period of flow measurement (post-1970). The peak instantaneous discharge of record of 818 m^3/s at the gaging station on the Sagavanirktok River was the result of a summer storm on August 19, 1974. That value compares with a maximum evident flood discharge of 1750 m^3/s measured by Childers and Jones (1975, table 1) at a site on the Sagavanirktok River 7.2 km upstream from the Lupine River confluence and a short distance downstream from site S2. Childers, Sloan, and Meckel (1973, p. 4) noted that the major flooding observed on the Sagavanirktok River in July 1961 may have formed the high-water marks observed in 1972 at a level 2 to 3 feet above the tops of the main channel banks. This flooding was observed approximately 20 km upstream from site S4, downstream from the confluence with the Ivishak River.

The unusual, but not catastrophic, 1976 summer storm in the Atigun River watershed was seen to have caused substantial channel change. The flow records (fig. 2) show that summer storms appear to be common in the Sagavanirktok River watershed, of which the Atigun River drainage is a part. Similar summer storms, however, have not significantly affected the recorded runoff in streams draining only foothill or coastal-plain areas. Not only the effects of the rarer floods, but their probable origin—rainstorm runoff or breakup—apparently may vary with physiographic province.

The evidence from other parts of the Arctic is likewise sketchy. On Ellesmere Island in the Canadian Arctic, Cogley and McCann (1976, p. 109–110) reported extensive shifting caused by summer-storm runoff in a braided channel from precipitation having a recurrence interval of perhaps 10 to 20 years. The degree to which the course of the main channel was changed was not clear, however.

There is no reason to suppose that the larger arctic stream channels are different from those in temperate regions in which the channel form is attributed to events of moderate frequency rather than the rare, catastrophic event (Wolman and Miller, 1960, p. 67). This frequency concept is not equally applicable to all streams. The smaller the stream the more variable the flow and the rarer will be the flood that causes channel change by exceeding the threshold of sediment movement or, in the case of arctic streams, causing sufficient thawing to release sediment for transport.

An example of the latter case may be the small Happy Creek drainage that showed essentially no channel change in spite of the fact that the 1976 breakup flooding was the greatest recorded. At the same time, the Sagavanirktok River at upstream sites showed significant amounts of change with levels of breakup flooding that were only moderate, based on the short period of record. That in the Arctic the events of moderate frequency—the bankfull floods of other climates—in fact correspond to the breakup floods has been a common suggestion. The results of this study confirm this suggestion for at least the larger streams.

LONG-TERM EROSION RATES

It is interesting to examine long-term erosion rates in light of the existing flow records. Brice (1971, p. 36–39) compared aerial photographs representing the period 1949–69 for 9.1 km of the Sagavanirktok River in the braided channel between sites S3 and S4. The flooding seen downstream in July 1961 would have affected these reaches during that period in the event the flow did not originate in the drainage basin of the Ivishak River. The photographs reveal a maximum of 91 m of lateral erosion, equivalent to 4.6 m/yr. Less than 12 percent of the main vegetated bank was affected by erosion detectable at the scale of the photographs, however, and Brice (1971, p. 38) was able to conclude that in general the 1949 river course was "not very different" from the 1969 course.

Childers and Jones (1975, p. 8) detected no erosion from aerial photographs at their site between October 17, 1969, and August 24, 1974. A brief comparison by the writer of 1969 and 1974 photographs in a series of reaches upstream from the Lupine River in the sinuous and meandering section of the stream similarly showed that detectable erosion in that time span was rare. The maximum amount seen was 34 m, equivalent to 6.8 m/yr and comparable to the maximum of 5.0 m of lateral erosion observed on the ground during the 1976 breakup period.

Thus, over a 27-year period surprisingly little lateral erosion has been observed generally along the course of the Sagavanirktok River, although significant amounts have occurred locally. This same result was found in the Canadian Arctic, where few areas of rapid lateral erosion in the previous 16 to 20 years were seen by Lewis and McDonald (1973, p. 268) in aerial photographs of streams draining the north slope of the Yukon Territory. The Alaskan results suggest, but by no means establish, that most of the erosion has taken place in small increments during breakup flooding. What erosion has occurred apparently has been concentrated in specific reaches where texture or stratigraphy of the bank is favorable for thermo-erosional niching. Other reaches have remained relatively unchanged since 1969, and possibly since 1949.

The above maximum amounts of lateral erosion cannot be compared directly with those of streams in other climates because of the variability of those data (see Wolman and Leopold, 1957, p. 96 and table 4). Certainly, however, the long-term rates of change in the channels of the streams in the study area appear to be small in relation to those of unregulated stream channels in most nonpermafrost environments. Assuming that absolute rates of erosion are in fact generally less than those in comparable channels elsewhere, the difference may be due more fundamentally to the climatic and hydrologic character of the permafrost environment than to the presence of permafrost in channel banks.

HYPOTHESIZED UNIFORMITY OF EROSION RATES

This study generally lends validity to the idea that arctic stream erosion may be comparatively uniform on an annual basis (Gill, 1972, p. 130; Cooper and Hollingshead, 1973, p. 276). Factors discussed above that support this hypothesis include: (1) the demonstrated effect of at least partial permafrost control of breakup channel processes; (2) the probability that breakup flooding is the most important channel-forming process in most larger watersheds; (3) what little is known of long-term erosion rates in the study streams; and (4) the similarity of those long-term rates with the maximum erosion observed during the breakup in 1976.

Uniformity in annual rates could occur as a result of comparable amounts of thermo-erosional niching and slumping in response to a relatively uniform annual thermal input to banks during breakup floods. The concentration of about half the annual precipitation in the breakup runoff of most streams is a phenomenon tending to smooth the vagaries of individual storms. Also, the variability of the precipitation in arctic ice-free barrens is reportedly smaller than that of dry coastal areas elsewhere (Giovinetto, 1974, p. 20).

Further work could establish that an average erosion rate determined from aerial photographs may represent a rate that could more confidently be extrapolated than the rates observed in nonpermafrost areas. There are a variety of obvious factors, however, that will mitigate against such extrapolation: susceptibility to thermal niching may vary laterally with the texture and ice content of the flood-plain deposits; icings and, to a lesser extent, ice jams may divert flow and cause large amounts of local erosion; and normal channel shifts will change the loci of maximum erosion.

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