Esker characteristics in terms of glacier physics, Katahdin esker system, Maine

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ABSTRACT

The characteristics of large, subglacially formed eskers, such as the Katahdin system, are closely related to two special peculiarities of water-filled tunnels along the beds of ice sheets: (1) the water pressure approximates the weight of the overlying ice; and (2) in tunnels that descend and those that ascend less steeply than ~1.7 times the ice-surface gradient, the walls melt, producing a sharply arched tunnel cross section, whereas in those that ascend more steeply, they freeze, producing a wide, low one. The first peculiarity primarily governs the paths of these eskers. It causes the tunnels to follow the paths ordinary rivers would follow were the land tipped downglacier ~11 times the local ice-surface gradient. The paths therefore trend in the general direction of the former ice flow but tend to deviate so as to follow valleys and to cross divides at the lowest passes, as observed. Ice-surface gradients calculated from path deviations at two localities on the Katahdin esker system indicate relatively thin, sluggish ice the surface of which lay ~200 m below the summit of Mount Katahdin, in agreement with independent geologic evidence. The second peculiarity primarily governs the form, composition, and structure of these eskers. Strong melting causes a large inflow of basal ice and entrained debris to the tunnel and produces sharp-crested eskers of poorly sorted, poorly bedded sand, gravel, and boulders with lithologies like the adjacent till, whereas weaker melting produces multiple-crested ones of similar composition. Freezing precludes inflow and produces broad-crested eskers of fairly well-sorted, well-bedded, more water-worn, coarse sand with few large clasts. Ice-surface gradients calculated from transitions from the multiple-crested type to small areas of broad-crested type on the Katahdin system agree closely with those computed from the paths at nearby localities. An anomalously low gradient calculated from a transition to an area of broad-crested type approximately twice as wide and long as the probable ice depth apparently confirms that, as expected, the basal ice was supported by water pressure over most, if not all, of the width of the esker.

INTRODUCTION

Tree-like esker systems tens, and even hundreds, of kilometres long occur throughout the regions of Europe and North America formerly covered by the great Pleistocene ice sheets. They were, with few exceptions, constructed by subglacial streams and rivers flowing in ice-walled tunnels along the glacier bed. Their relationship to moraines and other glacial features show that they generally formed during the terminal stages of glaciation when the ice was relatively thin and sluggish. The eskers comprising them can be as much as thirty or more metres high and can readily be traced on topographic maps across quadrangle after quadrangle. They trend in the general direction of the former ice flow but tend to deviate so as to follow valleys and to cross divides at the lowest passes. In fact, according to Stone (1899, p. 35), this resemblance to long-abandoned railway embankments led some of the oldtimers in Maine to refer to them humorously as "Indian railroads"! These general characteristics have been well described, with many details and examples, by Flint (1971, p. 214–218) and by Embleton and King (1975, p. 467–484), among others.

Not only the paths but also, as this paper will show, other characteristics of eskers, such as their form, composition, structure, and, more erratically, size, tend to correlate with the topography over which they pass. Clearly, this correlation, as well as the characteristics themselves, must depend upon the flow of ice and water near the glacier bed and hence indirectly upon the profile of the ice surface. Although this dependence is complicated by such details as the nonlinear nature of ice flow and the wide fluctuations in water flow, a good understanding of it can be attained through rather simple considerations, which this paper explores, using examples from the Katahdin esker system of Maine.

KATAHDIN ESKER SYSTEM

The Katahdin system (Fig. 1) stretches 150 km from its sources near Mount Katahdin in the middle of Maine to its terminus at Pineo Ridge close to the coast. It has five major tributaries resembling an elongate river network and spanning a maximum width of only ~30 km. This system and some of the other systems in Maine terminate in great kame deltas, the most prominent of which, Pineo Ridge in Cherryfield and Columbia, ~65 km east-southeast of Bangor, rises >60 m above the adjacent plain.

The continuity of the Katahdin system, the general increase in esker size downstream in it (Stone, 1899, p. 415), the absence along it of other kame deltas or significant outwash plains, and the fact that it cuts back and forth across subsequent positions of the retreating ice edge in the Penobscot Valley indicate that the whole system formed simultaneously (except possibly the terminal few kilometres), rather than piecemeal beneath a retreating ice margin. The single apparent exception is Silsby Plain in Aurora, 35 km east of Bangor and 4 km south of Morrison Ponds. Although considered by Stone (1899, p. 109) to have been constructed by water from the Katahdin system during retreat of the ice, it was actually deposited by water from an outlet several kilometres to the northeast (Van Beever, 1971, p. 18).

At the time of esker formation, ~12,700 yr ago, the edge of the ice in eastern Maine occupied a relatively static position at the former shoreline just north of the present coast (Borns, 1973, p. 42; 1978, p. 104). The ice flowed generally east of south, with some convergence in the vicinity of Pineo Ridge, as indicated by erratic blocks, glacial striations, drumlin axes, groove patterns, and other evidence, and as reflected by the esker paths.

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The land then lay ~80 m lower, relative to the sea, than now (Thompson, 1980, p. 213, 216), and the moraines and deltas were partially worked by marine waves and currents. Then, beginning shortly after 12,700 yr ago (H. W. Borns, Jr., 1983, personal commun.), the ice began its final retreat. In the low-lying areas, the sea followed it inland and spread a mantle of marine waves and currents. Then, beginning shortly after 12,700 yr ago (H. W. Borns, Jr., 1983, personal commun.), the ice began its final retreat. In the low-lying areas, the sea followed it inland and spread a mantle of shell-bearing clay over everything, including the eskers. This clay was named the "Presumpscot Formation" by Bloom (1960, p. 55; as cited in 1963, p. 865). In these areas, not unexpectedly, the retreat was extremely fast. Along the Penobscot Valley, for example, it amounted to 100 km in only a few hundred years, according to data given by Stuiver and Borns (1975, p. 100). In the higher areas, where the retreat was much slower, substantial remnants of ice probably survived another one to two thousand years (Borns and Hughes, 1977, p. 205).

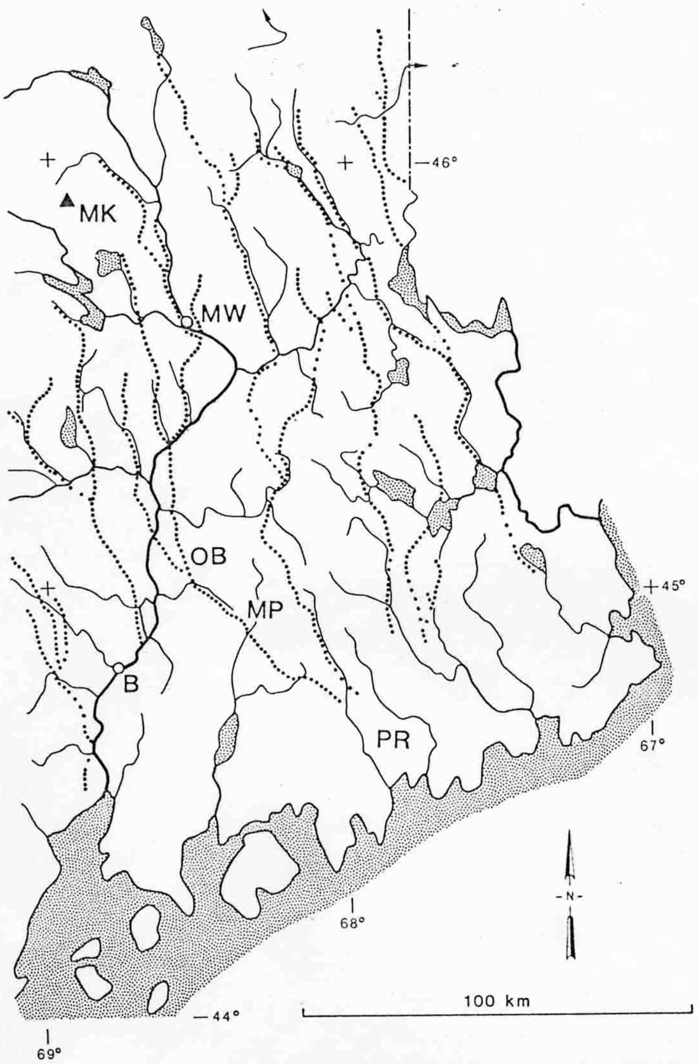

Figure 1. Major eskers of eastern Maine. Eskers (dotted) trend across present drainage (solid). Katahdin system heads in eastern foothills of Mount Katahdin (MK, elevation 1,600 m), passes Medway (MW), Otter Brook (OB), and Morrison Ponds (MP), and ends near Pineo Ridge (PR). Bangor (B) serves as convenient reference point. Based upon data from Stone (1899), Leavitt and Perkins (1935), J. B. Thompson, Jr. (1972, personal commun.), and observations by the author. Base map after U.S. Geological Survey Maine topographic sheet.

The esker systems trend diagonally across the rather deranged modern drainage (Fig. 1). The Katahdin system crosses passes ranging up to 100 m higher than the intervening valleys (Fig. 2). This is not exceptional (Flint, 1971, p. 215). There are many similar cases elsewhere in Maine and in the Canadian shield and northern Europe. Indeed, one esker system west of Hudson's Bay is >500 km long and ascends the regional slope to (Flint, 1971, p. 215). There are many similar cases elsewhere in Maine and in the Canadian shield and northern Europe. Indeed, one esker system west of Hudson's Bay is >500 km long and ascends the regional slope to an elevation 275 m higher than its headwaters (Wilson, 1939, p. 120, 127).

The Katahdin system, like other large eskers (Flint, 1971, p. 215), is commonly, though not always, discontinuous where it crosses passes (Stone, 1899, p. 413). The esker path in the gaps is delineated by a swath partially or wholly stripped of its till cover (Stone, 1899, p. 430). A good example is the pass crossed by the Katahdin system near Morrison Ponds, 35 km east-northeast of Bangor, where the esker path is characterized not only by bare bedrock ledges but also by scattered, rounded boulders and cobbles left by the former subglacial stream (Stone, 1899, p. 108).

In form, the large eskers of eastern Maine fall into three general types, which may be called *sharp-crested, multiple-crested,* and *broad-crested.* The multiple-crested type is also called "reticulate," especially in the older literature (such as Stone, 1899). The three types form a discrete series not only in form but also in situation, composition, and structure.

Sharp-crested eskers are by far the dominant type. They have steep sides and a single, sharp crest, the longitudinal profile of which is generally broadly undulatory although in places it may be prominently peaked. They are typically ~20 m high and 150 m wide but range from 3 to 50 m in height. They are situated in nearly level or gently descending reaches of esker paths. They typically consist of poorly to moderately sorted sand, gravel, and boulders, a few of which can be as much as a metre or more in diameter, deposited in a relatively high-velocity environment. Ripples, cross beds, and other sedimentary structures are generally scarce or absent. Bedding is poor and discontinuous, with numerous angular unconformities, and in most localities tends to dip away from the axis of the esker, forming an "anticlinal" structure or, in some of the most pronounced peaks, even a "domal" one (Stone, 1899, p. 39).

Although the term "anticlinal" (or "arched") is commonly used, the beds do not actually form an anticline but merely tend to dip away from the axis of the esker, in places nearly as steeply as the esker surface. This is a commonly observed feature and has frequently, although not always (Sharp, 1953, p. 872; Flint, 1971, p. 215–216), been attributed to slumping toward the sides in response to loss of support when the enclosing ice melted (Flint, 1971, p. 215; Embleton and King, 1975, p. 484). In the large Maine eskers, there is no evidence of significant slumping of this sort. Instead, as Stone (1899, p. 40) observed, it appears that the bulk of the sediments were initially deposited at the crest of the esker and then moved forward and downward along the flanks.

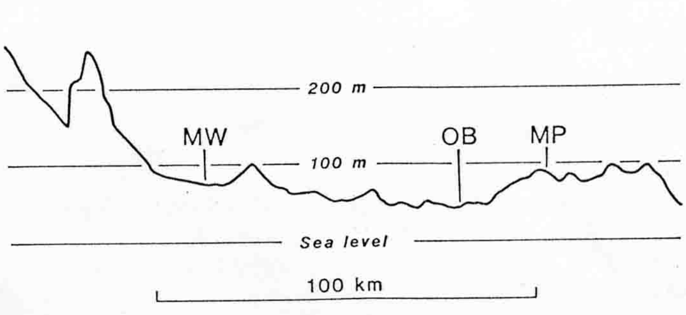

Figure 2. Longitudinal profile of main stem of Katahdin esker system. Locality symbols same as in Figure 1. Vertical exaggeration 200 ×.

The lithologies of the boulders and cobbles, and even of the pebbles and coarse sand, in sharp-crested eskers tend to be very similar to those in the adjacent till, which in turn tend to reflect the nature of the local bedrock (Stone, 1899, p. 431–432; Flint, 1971, p. 216). Because eskers are depositional features, the bulk of their sediment cannot come directly from the bedrock by fluvial erosion but instead must come indirectly by melting out of debris entrained in the basal ice by glacial erosion. This conclusion is supported not only by the observation of Stone (1899, p. 40) that the "anticlinal" structure is a primary feature but also by the observation of Trefethen and Trefethen (1944, p. 525–527) that fragments of the Hallowell Granite appear in the Kennebec Valley esker near Augusta at a point where its path, which lies wholly on schist, passes the nearest outcrops of the Granite at a distance of ~2 km.

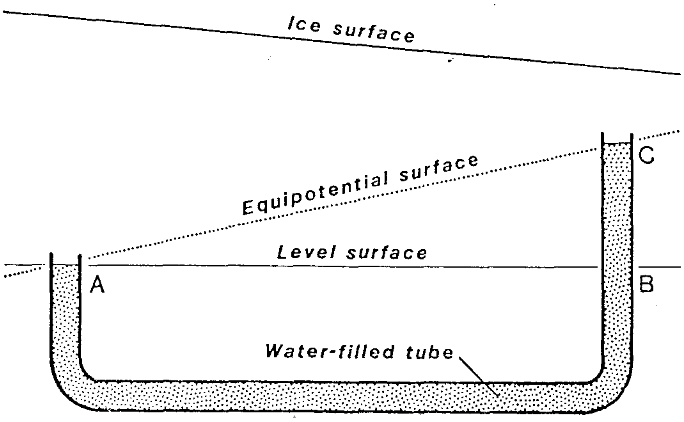

Figure 3. Gradient of hydraulic equipotential surfaces. Water by definition does not flow between points on equipotential surface; hence, weight of ice above A equals weight of water in column BC in hypothetical U-tube plus weight of ice above C. The gradient of equipotential surface is thus ~11 times gradient of ice surface.

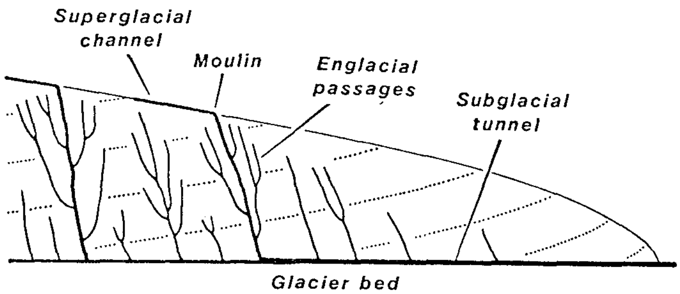

Figure 4. Equipotential surfaces (dotted) and drainage system in ice sheet, shown schematically for clarity. Water tends to follow lines of steepest descent in terms of the equipotentials. Vertical exaggeration ~50 to 100 ×; hence, actual equipotential surfaces would be almost horizontal.

Multiple-crested eskers generally have two to five or more parallel sharp crests, more or less randomly cross-connected, which are somewhat lower and rounder than the typical sharp-crested type. They form complexes that are typically ~450 m wide and as much as 15 m high but can be as much as 2 km wide. They are situated in gently ascending reaches of esker paths but otherwise are similar to the sharp-crested type. An excellent example is The Whalesback in Aurora, ~40 km east of Bangor.

Broad-crested eskers, which are not easily distinguished from proglacial fans or deltas, have a single, broad, flat crest that slopes gently to the sides, although the edges can be steep. They are typically ~600 m wide and 10 m high but can be as much as 2 km wide, and, like fans and deltas, are favored sites for highways, towns, farms, and sand pits. They are situated in the most steeply ascending reaches of esker paths (Stone, 1899, p. 448). They consist mainly of moderately to well-sorted sand with very few (or no) cobbles and boulders, which are more water-worn than in the other types (Stone, 1899, p. 441) and were deposited in a relatively moderate-velocity environment. Climbing ripples are common, many in very coarse sand, but cross beds, some of which may face upstream, are relatively rare. Bedding is good, is massive in places, and is apparently continuous over much or all of the breadth of the esker. It has fewer clear unconformities and is generally broadly "anticlinal."

In summary, the three esker types display distinctive assemblages of characteristics that form a discrete series in which the multiple-crested type is intermediate but more closely resembles the broad-crested type in situation and the sharp-crested one in form, composition, and structure. The purpose of this paper is to explain these characteristics and their interrelationships in terms of modern glacier physics, specifically, the principles of water movement under glaciers.

WATER UNDER GLACIERS

Water-filled tunnels along the beds of ice sheets resemble phreatic caverns in limestone, but they have two special peculiarities that govern the characteristics of the eskers that develop in them.

The first peculiarity is that the water pressure in such tunnels is controlled primarily by the weight of the overlying ice (Shreve, 1972). Ice is so deformable compared to limestone that tunnels more than ~100 m beneath the surface can significantly expand or contract in at most a few days in response to an increase or decrease in water pressure. The expansion or contraction lowers or raises the pressure in the tunnel by increasing or decreasing its carrying capacity until the internal and external pressures balance. Thus, the water-filled tunnels adjust in size so as to transmit exactly the amount of water supplied and to make the water pressure the same as the glaciostatic pressure in the surrounding ice. This is true regardless of whether parts of the glacier are subfreezing or where the water originates (although normally all but a tiny fraction of it comes from surface melting).

An important corollary of this adjustment is that water completely fills subglacial tunnels except locally, where the overlying ice is thin, or temporarily, when the water supply suddenly drops.

The direction of flow, and hence the path of a subglacial tunnel, follows the line of steepest descent in terms of hydraulic equipotential surfaces in the ice. By definition, water will not flow to points that lie on the same hydraulic equipotential surface as the starting point. Such a surface is therefore exactly analogous to a level surface in ordinary experience. If a hypothetical water-filled U tube connects two points on a hydraulic equipotential surface in an ice sheet with small surface gradient (Fig. 3), the weight of water and ice above a given level in the two arms of the tube must balance, inasmuch as the water pressure equals the glaciostatic pressure at the two ends. It follows that the hydraulic equipotentials dip upglacier with a gradient that is ~11 times as great as the downglacier gradient of the ice surface (Shreve, 1972, p. 211), as sketched in Figure 4. Thus, the flow of water under ice sheets is governed by the topography of the glacier bed and of the ice surface.

Distortions can occur, however, as a result of pressures caused by local deviations of flow (Fig. 5). The equipotentials are dimpled upward, for example, where the ice flows toward englacial passages or subglacial

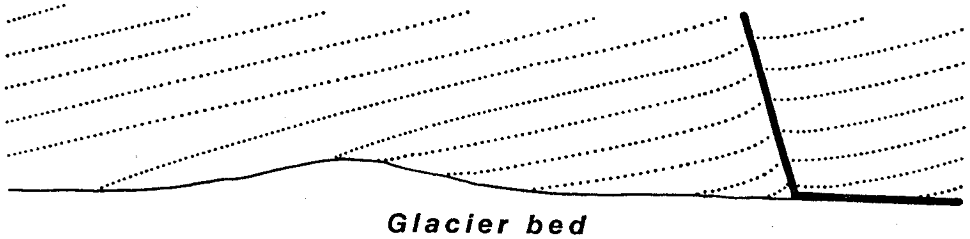

Figure 5. Distortion of equipotential surfaces by pressures due to local deviations of flow over subglacial ridge (left) and near englacial passage (right), shown schematically.

tunnels to balance melting of their walls. The resulting inclination of flow paths toward tunnels enables the larger tunnels to capture nearby smaller ones and thus develop the tree-like patterns of the larger esker systems. Similarly, the equipotentials are depressed in the regions of elevated pressure over the upglacier slopes of hills on the glacier bed. The resulting closer spacing of the equipotentials at crests causes a local increase in the transporting capacity of subglacial streams crossing them and thus produces the gaps in eskers at passes.

The second peculiarity of water-filled tunnels along the beds of ice sheets is that the melting of the tunnel walls, which is almost entirely due to viscous heating of the flowing water, tends to enlarge the tunnel, causing the water pressure to drop and inducing inflow of the surrounding ice. The rate of melting, however, is governed by the gradient of the tunnel (Shreve, 1972, p. 208) and can even be negative, that is, freezing can occur. In a level tunnel, the pressure decreases downstream because of the decrease in thickness of the overlying ice. The temperature of the walls, and hence of the water, therefore increases downstream, because the melting temperature of ice increases with decreasing pressure. In level tunnels, ~30% of the viscous heating is needed just to warm the water to the increased melting temperature, so that only 70% goes into melting the walls (Röthlisberger, 1972, p. 179). In tunnels ascending over subdued topography with gradient ~1.7 times the surface gradient of the ice, all of the heat is used to warm the water, and no melting occurs. In steeper tunnels, water freezes onto the walls, so that ice flows outward from the tunnel rather than inward toward it. The general relationship is plotted in Figure 6.

Where melting occurs, the tunnel shape is sharply arched, because the viscous heating, and hence the melting rate, is greatest where the floor-to-ceiling height, and hence the water flow, is greatest. Conversely, where freezing occurs, the shape is wide and low, because the necessary warming of the water, and hence the freezing of the walls, is greatest where the height is greatest. The first peculiarity governs primarily the paths of eskers, whereas the second governs primarily their form, composition, and structure.

ESKER PATHS

Esker paths follow routes perpendicular to the traces of the equipotential surfaces on the glacier bed, just as surface water follows routes perpendicular to topographic contours, which are the traces of level surfaces on the surface of the Earth. The esker paths will thus follow the valleys on equipotential contour maps of the glacier bed. For gentle bed topography and low ice-surface gradients, such maps can be constructed to useful accuracy from the geometry of the ice surface and the glacier bed by means of the principle that the equipotentials dip upglacier with gradient ~11 times that of the ice surface. Figure 7 gives details of a graphical method.

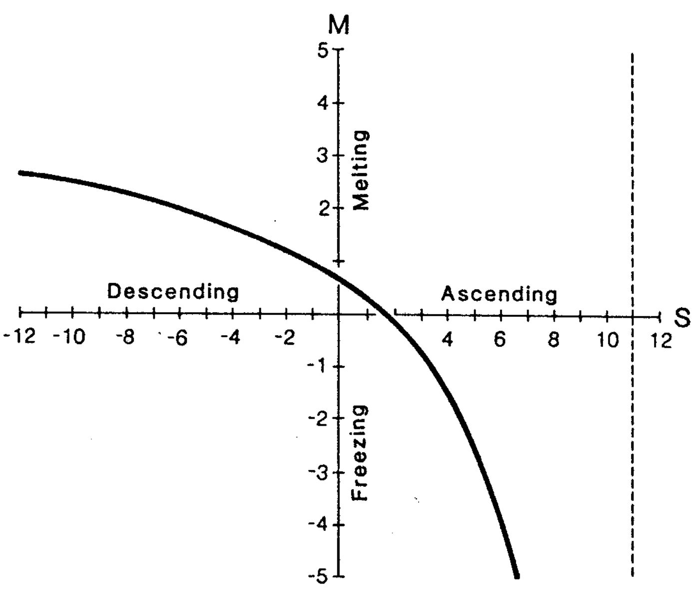

Figure 6. Rate of wall melting as function of gradient of subglacial tunnel. M is ratio of wall-melting rate to viscous-heating rate; S is negative of ratio of glacier-bed gradient to ice-surface gradient in direction of esker path. Relationship is approximation inaccurate for gradients greater than ~500 m km^{-1}, far steeper than any esker path.

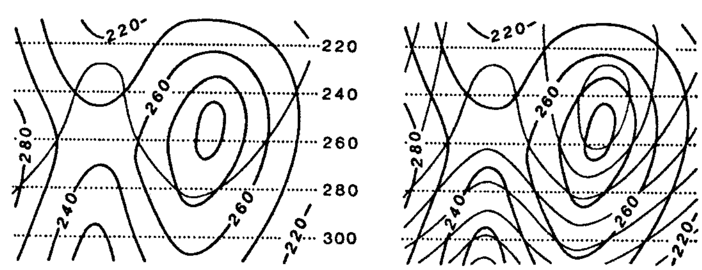

Figure 7. Graphical construction of equipotential contour map. To construct single contour (left), superpose topographic map of a single equipotential surface (dotted) with same scale and contour interval on topographic map of the glacier bed (heavy solid) and draw locus of points which are at same elevation on both maps (light solid). The horizontal spacing of the contours on the equipotential surface will be the contour interval divided by 11 times the ice-surface gradient. For example, if the scale and contour interval are 1:62,500 and 20 ft (6.1 m), as for much of Maine, and the ice-surface gradient is 6.8 m km^{-1}, the spacing will be 20/(11 × 0.0068) = 270 ft (81 m) on the ground, which corresponds to 0.051 in. (1.3 mm) on the map. To construct other contours (right), repeat process for equipotential surfaces spaced at equal intervals above and below the first. The most convenient spacing (used here) is the contour interval (or an integral multiple of it), because the contour lines representing the equipotential surfaces need only be mentally relabeled, not physically repositioned.

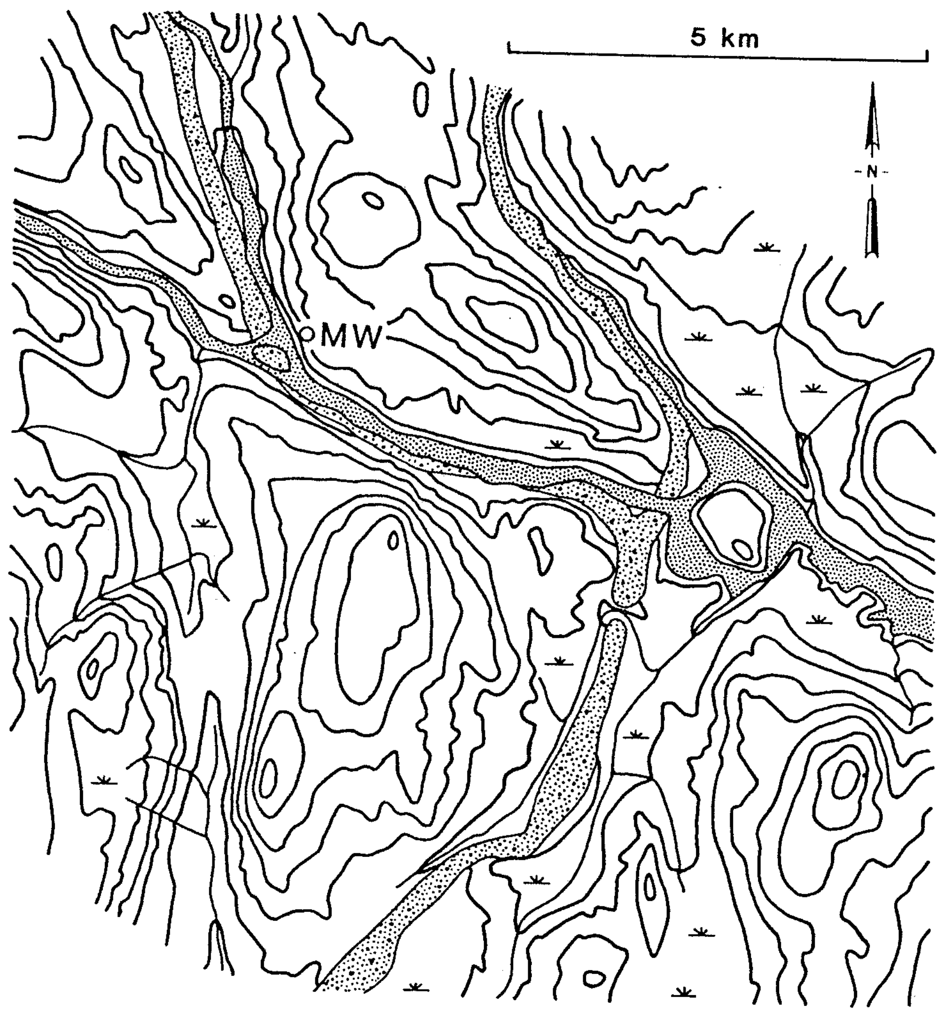

Figure 8. Topographic map of Medway area showing esker (mixed dots). Ice flowed ~S16°W. Contour interval is 40 ft (12.2 m). Center of town (MW) serves as convenient reference point. Based upon U.S. Geological Survey Millinocket and Mattawamkeag topographic quadrangle maps.

The equipotential contours will be like the ordinary contours obtained if all parts of the bed were tipped downglacier ~11 times the local ice-surface gradient. The esker paths will thus trend in the general direction of ice flow, which tends to parallel the ice-surface gradient, but will deviate so as to follow valleys and to cross divides at the lowest passes, just as observed. They will also tend to follow troughs in the ice surface, such as form in the lee of nunataks. In general, the steeper the ice surface, the less the esker paths will be influenced by bed topography.

The catch in making an actual equipotential contour map is, of course, that the gradient of the ice surface is usually unknown. In suitable cases, however, it is possible to reverse the process, that is, to approximate the gradient from the known esker path by systematically trying various gradients until one is found that best accounts for the path.

The Katahdin esker system provides an excellent example of such a case where it leaves the Penobscot River near Medway, ~90 km north of Bangor. As shown in Figure 8, the esker follows the south side of the river valley eastward almost perpendicular to the general ice-flow direction for 5 km, then turns sharply southward and cuts across the high ground through a gap at almost exactly the same elevation as an adjacent one to the west that it bypassed. The till and other superficial deposits are generally only a metre or so thick, postglacial erosion has been inconsequential, and, as throughout much of eastern Maine, topographic slopes are gentle. It should thus be possible to bracket the former ice-surface gradient in the area fairly closely.

Figure 9 shows the equipotential contours, uncorrected for local distortion, for an ice-surface gradient at Medway of 6.8 m km^{-1}. This is the local gradient given by the commonly employed square-root ice-surface profile (Nye, 1952, p. 529) based on the assumptions that the basal shear stress is uniform, that it has a magnitude of 100 kPa (1.00 bar), that the bed at the scale of the ice sheet is essentially a level plane, and that the terminus was at Pineo Ridge, 120 km along the line of flow to the southeast. In terms of the equipotentials, the river valley is only a less steeply inclined shelf on a sloping surface. It is clear why the eastern branch of the esker turned south instead of following the river valley, but the western branch should have done the same. The ice-surface gradient must therefore have been less than assumed.

Figures 10 and 11 show the contours for gradients of 4.8 and 3.4 m km^{-1}, which correspond to basal shear stresses of 50 and 25 kPa (0.50 and 0.25 bar). The smaller gradient appears to be too small, whereas the larger one seems about right. It corresponds to an ice depth at Medway of 1,150 m. Projection of the profile 35 km northward to the latitude of Mount Katahdin puts the ice surface ~200 m below the elevation of the summit when the eskers were forming, in agreement with geologic evidence from elsewhere in the highlands of northwestern Maine (H. W. Borns, Jr., 1983, personal commun.). Shreve (1985) gives a more detailed calculation of the ice-surface profile and basal shear stress.

The basal shear stress of 50 kPa (0.50 bar) tends to agree with the geologic evidence for relatively thin, sluggish ice at the time of esker formation. It implies thicknesses only two-thirds, and velocities only one-eighth, those of modern, active ice sheets with basal shear stresses of

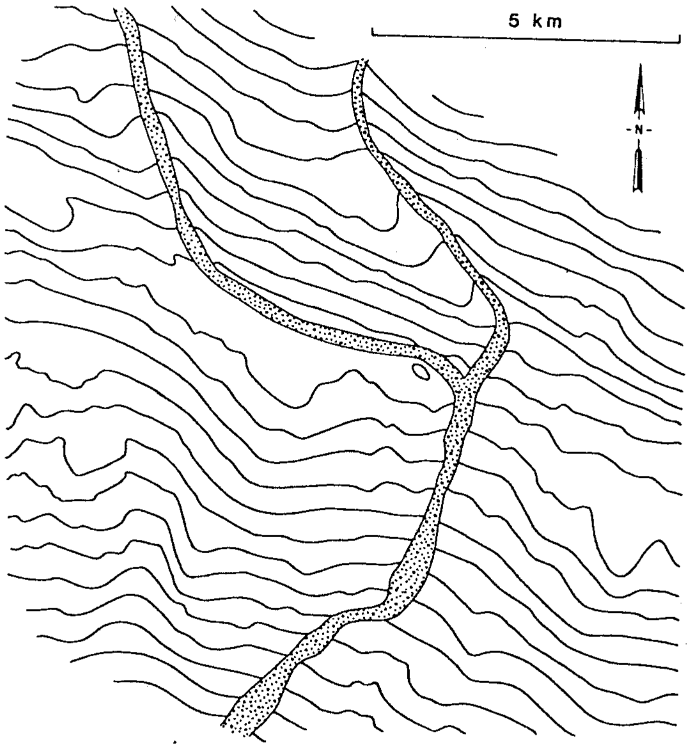

Figure 9. Equipotential-contour map of Medway area constructed for ice-surface gradient of 6.8 m km^{-1}, which corresponds to uniform basal shear stress of 100 kPa (1.00 bar). Contour interval is arbitrary.

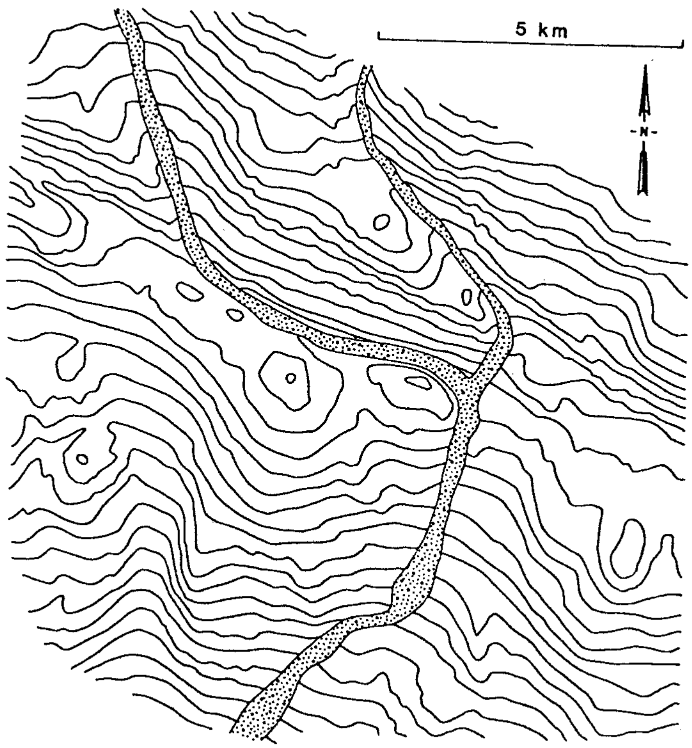

Figure 10. Equipotential-contour map of Medway area constructed for ice-surface gradient of 4.8 m km^{-1}, which corresponds to uniform basal shear stress of 50 kPa (0.50 bar). Contour interval differs from that in Figures 9 and 11.

100 kPa (1.00 bar). The ice was not "stagnant, " however. The existence of the esker systems demonstrates that it sloped smoothly to its margin 120 km away near the coast. It was therefore flowing forward everywhere, although much more slowly than are typical present-day ice sheets.

A puzzling feature on the equipotential contour map is the sharp kink in the esker where it crosses the contours diagonally ~4 km south of the river valley. It may be due to the pressure shadow in the lee of the broad hill immediately to the northwest, which would tend to bow the contours northward. This feature is not an isolated anomaly. Other clear examples on the Katahdin system are in the lee of Whetstone Mountain, 25 km northwest of Medway, and of Hazeltine Ridge, 30 km southwest of it.

A disadvantage of the equipotential-contour-map method of inferring the ice-surface gradient is that it is extremely laborious. Moreover, most of the labor produces information that does nothing to constrain the gradient. In the Medway case, for example, nearly all of the crucial information comes from the small area where the esker turns east along the river instead of continuing south through the bypassed gap. In this area, which is 1.3 km S30°E of the center of the town, the ground surface is inclined obliquely to the ice-flow direction in such a way as to make the esker path sensitive to small differences in ice-surface gradient. The ground surface slopes downward N31°E with gradient 40 m km^{-1}, the ice-flow direction was S16°W, and the esker direction is S43°E. Equating the apparent dip (formula given by Billings, 1972, p. 522) of the ground surface to that of the equipotential surfaces in the direction of the equipotential contours (which for small ground-surface gradients can be taken perpendicular to the esker direction), then solving for the tangent of the true dip of the equipotential surfaces, and finally dividing by 11 gives 4.1 m km^{-1} for the ice-surface gradient.

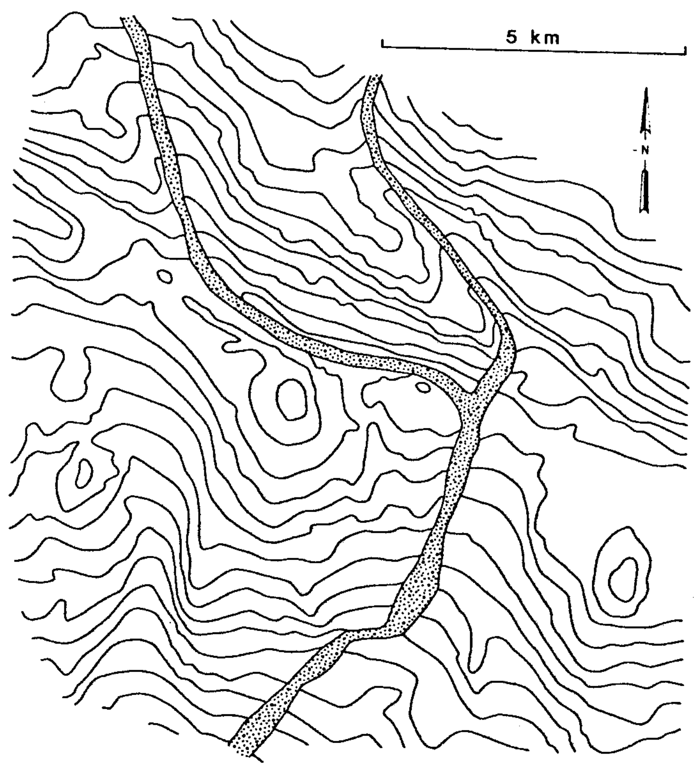

Figure 11. Equipotential-contour map of Medway area constructed for ice-surface gradient of 3.4 m km^{-1}, which corresponds to uniform basal shear stress of 25 kPa (0.25 bar). Contour interval differs from that in Figures 9 and 10.

ESKER FORM, COMPOSITION, AND STRUCTURE

The rate of melting or freezing of the ice walls strongly influences the form, composition, and structure of eskers, because it controls both the shape of the subglacial tunnel and the influx of rock debris.

Sharp-crested eskers result from the strong melting associated with near-level or descending reaches of esker paths. The wall melting, and hence the water flow, must be greatest near the crest, because the bulk of the debris appears to be initially deposited there and then moved forward and downward along the flanks. At first sight, this seems anomalous, because the water will tend to seek the lower paths available along the flanks of the esker. On the other hand, the debris influx to a tunnel along one of the flanks is highly asymmetric, because the ice arriving at the lower side is more heavily debris-laden. Moreover, debris arriving higher up will tend to be washed to the lower side. The accumulated debris protects that side from melting, causing preferential melting of the upper side and hence migration of the tunnel toward the crest. Where strong melting occurs, the subglacial stream may therefore be somewhat unstable, spending most of its time at the crest but repeatedly "slipping" down one flank or the other (or both at once) and then migrating back to the top, in the process generating the characteristic discontinuous bedding, angular unconformities, and "anticlinal" structure.

The inward flow of ice caused by the melting carries entrained debris toward the tunnel from the sides. This explains why the cobbles and boulders in sharp-crested eskers closely resemble the adjacent till and why they include lithologies from sources off the esker path. Because the inward flow is fastest near the tunnel, the debris follows sweeping trajectories that begin almost parallel to the tunnel and curve inward to end almost perpendicular to it. The shape and extent of the converging trajectories are governed by the rate of melting. Efforts to prospect for bedrock mineral deposits by means of samples from eskers must therefore take account of esker size and, as will be seen, type.

Multiple-crested eskers result from the lower rate of melting of the tunnel walls associated with gently ascending reaches of esker paths where the gradient approaches 1.7 times that of the ice surface. The tendency of the subglacial stream to migrate upward is weaker than in nearly level or descending reaches, whereas the tendency to "slip" down the flanks is undiminished. The result is instability at lower crest heights and a tendency to construct new ridge segments somewhat at random along the flanks of the existing esker. A corollary is that sharp-crested eskers will eventually become multiple-crested if allowed to grow high enough, a possible example being The Whalesback in Aurora, ~40 km east of Bangor. The point of transition from the one form to the other thus has no special glaciological significance.

Broad-crested eskers result from the freezing of the tunnel walls associated with steeply ascending reaches of esker paths where the gradient exceeds ~1.7 times that of the ice surface. The esker shape is broad and flat, the water velocity is moderate, and the sediment bedding is more continuous because the tunnel is wide, low, and stable. Finally, the sediment is finer and rounder, because in the absence of any influx by wall melting, it all comes from upstream.

The continuity of the bedding indicates that the subglacial stream covers much of the width of broad-crested eskers, if not all of it, simultaneously. This means that the water pressure supports the weight of the ice, and the basal shear stress is effectively zero, over a width typically comparable to, and in some cases double and even triple, the probable ice depth. This results in a stream of faster-moving ice marked by a shallow longitudinal trough of lower gradient in the ice surface. The corresponding broad ridge in the equipotential surfaces produces the broad convexity of the esker crest, inasmuch as the water flow is parallel to the esker axis while necessarily perpendicular to the equipotential contours. In other words, on an equipotential contour map, the convex esker surface forms the bottom of a flat-bottomed valley, even where the ground surface is level. The tendency of the subglacial stream to spread is limited by the valley walls. Like an ordinary valley fill, therefore, the esker widens only as it thickens (at least until its width is much greater than the ice depth), thereby producing the broadly "anticlinal" internal structure.

The spreading of the subglacial stream, although not of the esker, is also limited by the tendency for channels to develop in the sediment. Ultimately, when the esker grows big enough, the stream must wander from side to side over it, like a braided river on outwash.

The convergent flow of ice toward the upglacier end of the longitudinal trough in the ice surface causes an upward dimpling of the equipotential surfaces, just as does the inflow of ice toward a subglacial tunnel in response to wall melting. This lowers the gradient of the equipotential surfaces and enables the point of transition in esker type, and with it the relatively steep upglacier end of the trough, to migrate headward until stopped by a sufficiently unfavorable gradient in the esker path.

The transition from multiple-crested to broad-crested provides a means of approximating the ice-surface gradient independently of the methods based on the esker path, provided the broad-crested portion has a small width or length in comparison to the ice depth. In the Medway area, the topographic map indicates such a transition at the head of Little Pattagumpus Stream, ~7 km south of the center of town. At that point, the esker path climbs 7.1 m km^{-1}, which, when divided by 1.7, gives an apparent ice-surface gradient of 4.2 m km^{-1} in the direction of the esker, S42°W. When corrected by means of the apparent-dip formula to the ice-flow direction, S16°W, this gives a true gradient of 4.6 m km^{-1}, in moderate agreement with the value obtained by the oblique-path method. Unfortunately, this locality is close to the puzzling kink in the esker. Moreover, owing to poor accessibility, it was not field checked.

A similar comparison can be made on the Katahdin system at a transition located 1.0 km north of where the esker crosses Otter Brook, ~35 km northeast of Bangor and 60 km from the former glacier terminus. At this locality, which was field checked, the ground surface slopes downward N34°E with gradient 15 m km^{-1}, the ice-flow direction was S14°E, and the esker direction is S30°E. With these data, the oblique-path

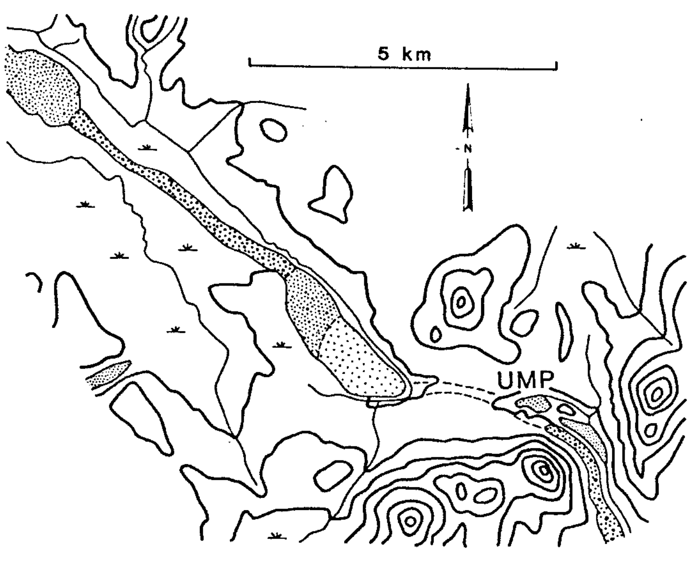

Figure 12. Topographic map of Morrison Ponds area showing esker with sharp-crested, multiple-crested, and broad-crested reaches (coarse, mixed, and fine dots). Esker is missing in pass (dashes). Contour interval is 100 ft (30.5 m). Upper Morrison Pond (UMP) serves as convenient reference point. Based upon U.S. Geological Survey Great Pond topographic quadrangle map.

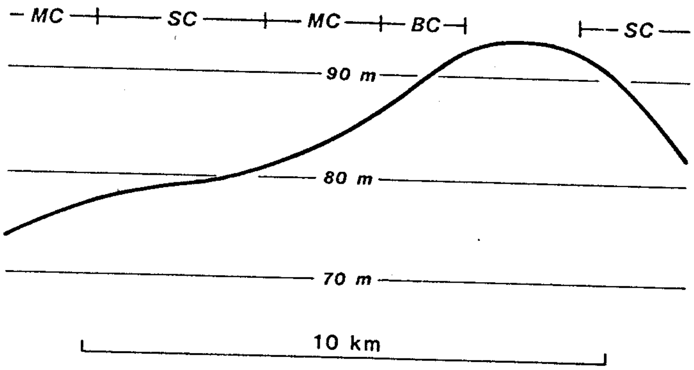

Figure 13. Longitudinal profile of esker path in Morrison Ponds area showing correlation of esker type with gradient. As gradient steepens, type changes from sharp-crested (*SC*) to multiple-crested (*MC*) to broad-crested (*BC*). Vertical exaggeration 200 ×.

method gives 4.4 m km^{-1} for the ice-surface gradient, and the transition method gives 4.0 m km^{-1}, again in good agreement. Indeed, in light of the uncertainty in the data, especially the orientation of the ground surface, this close agreement must be regarded as at least somewhat fortuitous. Nevertheless, the consistency of the two methods in the two areas does tend to support the underlying model of esker mechanics.

A reach of broad-crested esker almost 1 km wide and >1.5 km long, by far the widest and longest in the Katahdin system, is located 3 km northeast of upper Morrison Pond, 35 km northeast of Bangor, and 35 km from the former terminus. In this locality, which was also field checked, the esker ascends toward a low pass through a prominent range of hills that rises 100 to 200 m above the general level (Fig. 12). The longitudinal profile, plotted in Figure 13, clearly shows the close correlation of esker type with gradient. The gradient at the transition to broad-crested is ~3.2 m km^{-1}, which implies an ice-surface gradient of only 1.9 m km^{-1}. This is less than half the values found at Medway and Otter Brook, whereas it should be greater, inasmuch as it is closer to the terminus. Although, as usual, the measured gradient is subject to considerable uncertainty, the error definitely cannot be great enough to account for the discrepancy. Considering that for a uniform basal shear stress of 50 kPa (0.50 bar) the ice thickness would have been only ~600 m, this appears to be a striking confirmation of the headward migration of the transition point into a region of unfavorable esker-path gradient.

ESKER SIZE

The size of an esker at any given point on its path is proportional to the cumulative volume of sediment delivered from the ice by wall melting and from upstream by fluvial transport less that carried away downstream. The sediments carried by the stream are deposited where the transporting capacity diminishes. In general, this occurs where the tunnel turns less steeply downward or more steeply upward, where the spacing of equipotentials increases, or where the tunnel changes to a less efficient shape, as at transitions from melting to freezing of the walls. The rate of sediment delivery from the ice, on the other hand, is proportional to the rate of melting of the walls and to the concentration of debris in the basal ice. The melting rate, in turn, is proportional to the water discharge and, in a level tunnel, to the ice-surface gradient (Shreve, 1972, p. 208, equation 17). In inclined tunnels, the melting rate decreases or increases, but not proportionally, the more steeply the tunnel ascends or descends, as shown in Figure 6.

Esker size therefore depends on the duration of accumulation and on a complex variety of other factors. The downglacier increase in water discharge and ice-surface gradient, and hence of available sediment, accounts for the general downstream increase in size in large eskers such as the Katahdin system. Where the accumulating sediment is delivered entirely by the stream, as is true of all broad-crested eskers, the size will be greatest where the longitudinal profile is concave upward, the ice-surface gradient becomes more gentle downglacier, and the esker type changes to broad-crested. Where the sediment is delivered entirely by the ice and exceeds the competence of the subglacial stream, as is approximately true of some sharp-crested eskers, the size will be greatest where the debris concentration is greatest, the water discharge is highest, the ice-surface gradient is steepest, and the tunnel gradient is downward. In the vast majority of cases, however, neither process dominates, and both sets of factors influence the size, which explains its wide local variability.

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