

A Braided River System in a Glacial Environment, the Copper River, Alaska

by John Wooster

INTRODUCTION

Climate and geology are the primary forces shaping the Earth's landscape. Climate and geologic processes combine to control water discharge, sediment supply, and channel slope, which dictate a river channel's pattern. Glacial processes typically produce high discharges, large volumes of sediment, and steep channel slopes. Rivers respond to these processes by developing a braided river pattern. Braided rivers contain multiple separated by bars or islands. Braided channels are variable, dynamic systems with high rates of fluvial activity and channel adjustment from erosional and depositional processes. Although the threshold between meandering (sinuous, single channel river pattern) and braiding is not clearly understood, Knighton (1998) identifies three factors probably necessary for braiding to occur: an abundant bedload supply (portion of a river's sediment load supported by the channel bed), erodible banks, and high stream power (an expression for the potential energy expenditure for a given river channel length).

The Copper River is located in south-central Alaska and drains into the Gulf of Alaska at the southeast edge of Prince William Sound. The Copper River is a prime example of a braided river system found in high latitude, glacially dominated environments (figure 1). The Copper River basin is heavily glaciated (approximately 18%), and the system produces high water discharges and suspended sediment loads relative to other major Alaskan basins (Brabets 1997). Even though the Copper River is potentially an unstable environ, the watershed provides habitat for a diverse and prolific Pacific Salmonid population [refer to (Koenig, 2002) in this volume for more detail].



Figure 1: View of the Copper River and braidplain near the Copper River Delta (USFS 2000).

BRAIDED SYSTEMS

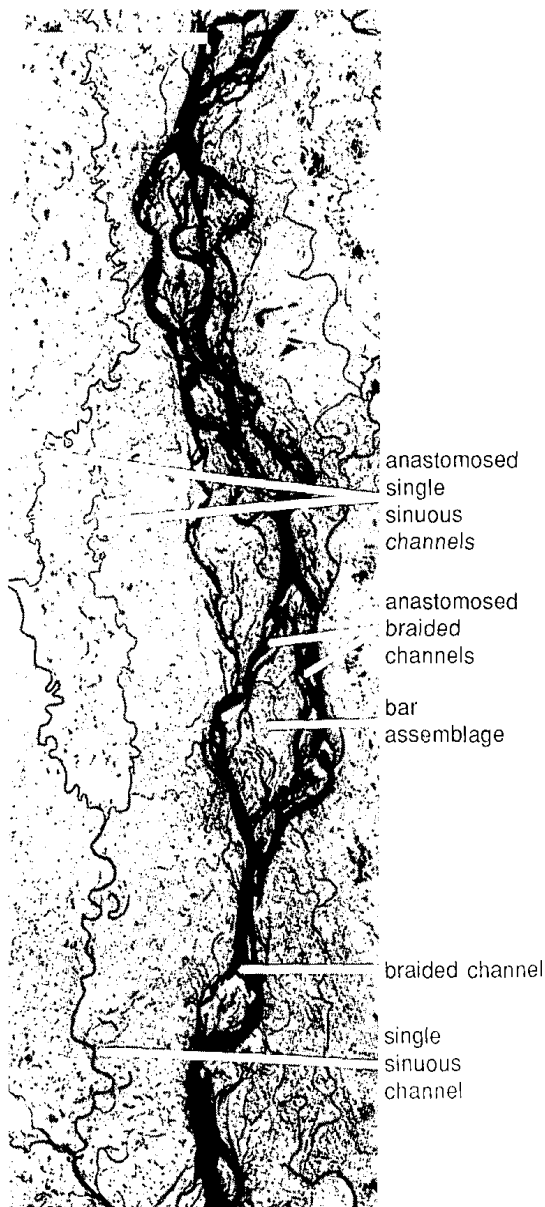


Figure 2: Examples of braided and anastomosed channels on the Brahmaputra River (Bridge 1993).

River channel patterns can be regarded as a continuum with three general characteristic morphologies: straight, meandering, and braided. The threshold between meandering and braiding represents a primary subdivision of channel patterns into single and multiple channel forms. Early definitions of braided rivers in the literature can be found in Leopold and Wolman (1957) who defined a braided river as “one which flows in two or more anastomosing channels around alluvial islands”, and Lane (1957) stated “a braiding stream is characterized by having a number of alluvial channels with bars or islands between meeting and dividing again, and presenting from the air the intertwining effect of a braid.” Schumm (1977) provided further clarification by defining braided channels as single-channel bedload rivers which at low water have islands of sediment or relatively permanent vegetated islands, in contrast to multiple channel rivers (anastomosing or distributive) in which each branch may have its own individual pattern. Braiding is typically referred as to the splitting of channels around bars or islands, which are contained within a dominant pair of floodplain banks. Anastomosing refers to a type of channel

splitting where the channel segments are divided by stable, often vegetated islands that are large relative to channel widths (figure 2). Anastomosed channels typically remain divided at bankfull flows (often referred to as the formative discharge that equals or barely exceeds the top of a river’s banks) and their width scale flow patterns behave independently of adjacent channels (Bridge 1993). The main characteristic of the braided reach is the repeated division and joining

of channels and their associated flow patterns, referred to as zones of confluence and diffuence. Ashmore (1991) defines braiding as a “distinct bifurcation of the flow and/or bed-load flux (sediment transport along the channel bed) around inactive portions of the channel bed, and need not necessarily involve exposure of an inactive bar above the water surface.”

Braided fluvial systems are abundant throughout upland and proglacial systems, and are common in both gravel and sand-bedded channels. Braided systems have a wide range in scale, varying from braidplain widths of tens of meters up to 20 km found in large alluvial systems such as the Brahmaputra River. Braided rivers are distinctive because of their high stream power (product of specific gravity of water, slope, and discharge) and subsequent high rates of erosion, deposition, and channel change compared to other river types. Typically braiding is associated with high values of valley slope, sediment supply, stream power, shear stress (product of specific gravity of water, slope, and water depth), width to depth ratio (channel width to average flow depth), and bedload transport rates (Ferguson 1993). The Copper River represents a classic proglacial braided system with relatively high discharges, sediment loads, and slope. Compared to the wealth of literature on meandering rivers, braided rivers have been relatively understudied. Braiding rivers pose a significant challenge to study due to the difficulty of measuring flow, sediment transport, and morphology in the rapidly shifting braided river environment. Several key issues in braided river environments require further attention: mechanisms of braid bar initiation, channel confluence and diffuence (channel splitting) dynamics, and the influence of flow stage and aggradational regime (periods where a system is undergoing net increase in sediment storage) upon depositional patterns (Bristow and Best 1993).

Theories on the causes of braiding in alluvial systems can be divided into three general categories. The first is a functional explanation relating braiding to a combination of externally imposed environmental factors such as discharge and sediment supply (Ashmore 1991). The second class of hypotheses as to braid formation is theoretical stability analyses of channel-scale bed forms (primarily bars) in two dimensional flow regimes (Parker 1976; Hayashi and Ozaki 1980). The final class focuses on the physical sedimentary processes and hydraulic conditions that accompany the onset of braiding.

External Factors

The bulk of theories and analyses attempting to explain the reasons for braiding in a river system involve establishing a correlative relationship between braiding and environmental characteristics such as watershed area, discharge, channel or valley slope, sinuosity, width to depth, sediment supply, sediment particle size, and bank resistance (Ashmore 1991). The empirical relationships often use regressions of known braided system parameters to form a lower boundary for braided rivers. The lower boundary thus delineates the meander/braided threshold. Generally, it is accepted that channel pattern is controlled by water and sediment supply, and in the short term by slope. Slope and water discharge can be quantified more readily and accurately than sediment supply, and usually form the bases for empirical meander/braided thresholds. The slope-discharge relationships typically work well when derived and applied in a regional context, but often break down when attempting to categorize a broad range of braided systems. Examples of regression relationships are given in textbox 1.

Textbox 1: Empirical thresholds for meandering versus braided channel patterns.

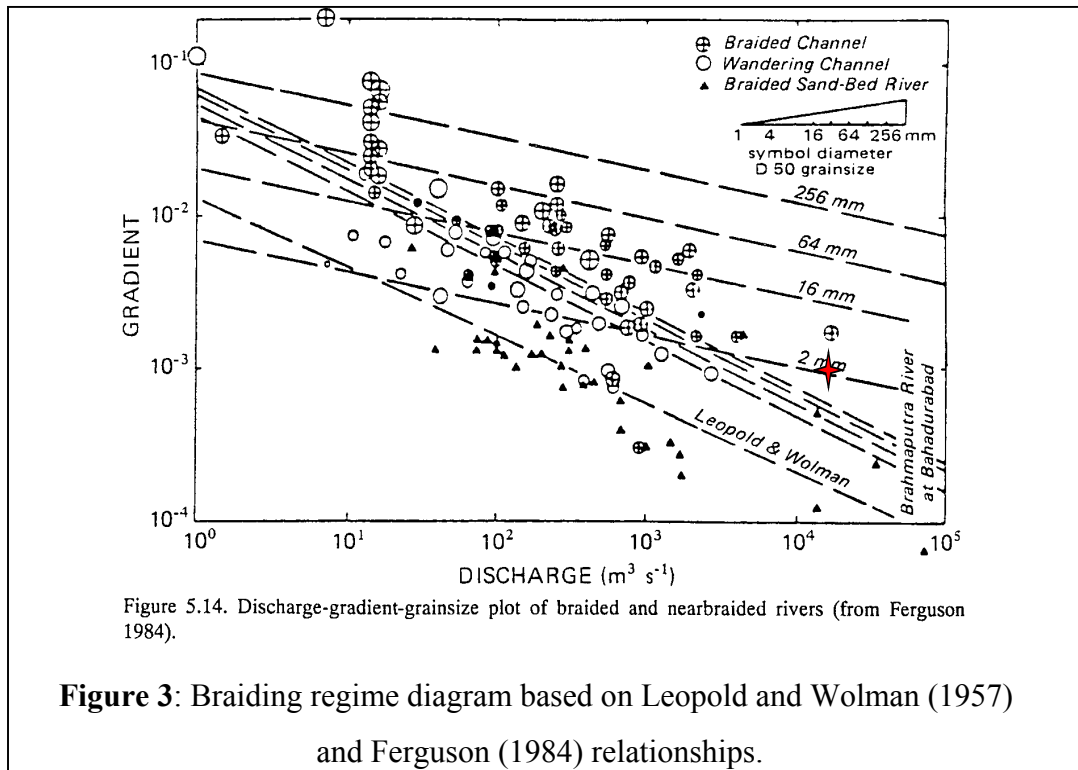
$$S = 0.0125Q_{bankfull}^{-0.44} \quad (\text{Leopold and Wolman, 1957}) \quad (1)$$

$$S = 0.042Q_{bankfull}^{-0.49}D_{50}^{0.09} \quad (\text{Ferguson, 1984}) \quad (2)$$

S = slope, Q = water discharge, D_{50} = median particle size

The above relationships define meandering/braided thresholds.

Slope and discharge have an inverse relationship. When the slope is greater than predicted for a given discharge the river falls into the braided regime, and slopes less than predicted in the meander regime. The same holds true for discharge (i.e., braided patterns are predicted for discharges greater than estimated for a given slope.) As either slope or discharge increase, the value of the other parameter necessary for braiding decreases. Figure 3 displays equation 1 and 2; the shallow lines are derived from Parker's (1976) stability analysis described below and the shallow lines represent equation 2 (with the same grain sizes as labeled on the shallow lines: 256, 64, 16, 2 mm.) The Copper River is plotted as a red star in figure 3.



High values of channel slope are often referred to as prerequisite for braiding. However, braided does occur with relatively low slopes in systems that have very high discharges. Thus a more appropriate prerequisite for braiding might be high stream power, which combines slope and discharge (Knighton 1998). The bed material size distribution also plays a major role in degree of braiding. As illustrated in figure 3 and equation 2, smaller grainsizes provide less boundary resistance and allow braiding to occur at lower slopes and discharges. Rivers with high discharge variability can have a greater propensity for braiding due to rapid fluctuations in discharge that typically produce high sediment supply and width to depth ratios. The extreme discharges lead to bank erosion and irregular, episodic bedload movement, which are important to braided channel formation (Knighton 1998). Systems with high discharge variability are common in proglacial systems and in the Copper River [refer to (De Paoli 2002) in this volume for more detail]. Discharge variability is not a prerequisite to braiding, as flume experiments have demonstrated braiding will form at constant discharges and braided reaches can be interspersed with meandering ones.

Another key component of braided systems is bank erodibility. Banks of readily erodible material are key sources of sediment as well as being necessary for channel widening, which is a

key component of braided reaches. A braided system must have sufficient power to erode its banks and achieve high bed mobility (Knighton 1998). The highly unstable bed and banks found in braided systems and the Copper River tend to compromise the quality of spawning and rearing habitat for anadromous fish [refer to (Wheaton 2002) in this volume for more detail]. Thus the Copper River serves primarily as a migration path for salmonids to access tributaries with more favorable habitat conditions [refer to (Koenig 2002) in this volume for more detail]. Vegetation also plays a significant role in determining a river's channel pattern. Vegetation can have a positive feedback for braiding by stabilizing cut banks and bar surfaces; however, vegetation can increase bank resistance and inhibit channel widening, which promotes a meandering channel pattern. Once established, the unstable, rapidly shifting braiding river environment serves as a major disturbance force in vegetation establishment and successional regimes [refer to (Trowbridge, 2002) in this volume for more detail].

Braiding Indices

In order to establish the degree to which a braided stream network has developed and to quantify how such a pattern changes through time, it is important to derive some measure of degree (index) of braiding. However, developing a consistent, repeatable braiding index proves to be a formidable task due to rapidly changing channel morphologies and difficulties of access and accurate measurement in the large number of stream channels typical of braided systems. Additionally, many hydraulic geometry properties of braided streams are extremely stage dependent and cannot be quantified in a repeatable method at different flows. Braiding indices generally fall into two categories: those that count the mean number of active channels or braid bars per channel transect, and those that consider the ratio of the sum of channel lengths in a reach to the total reach length (total sinuosity) (Bridge 1993). Bridge states that the first category is preferable because it relates braiding to the number of alternate bars or channels, which stability analyses have shown are a fundamental component of braid development. Total sinuosity can be an indicator of braiding, but is not necessarily a controlling variable in braid establishment.

Theoretical Stability Analysis

Theoretical stability analyses use perturbation techniques for equations of motion to analytically define values for meandering/braided thresholds. Theoretical stability analyses are all based on the inherent instability of flow and sediment transport. In a simplified view, the primary factor causing the instability is the phase difference between the shear stress gradient (a function of fluctuating water elevations) and the bed form gradient (elevations of the channel bed) (Hayashi and Ozaki 1980). Sediment transport and bed friction are necessary conditions for the instability (Parker 1976). One of the early pioneering stability analysis by Engelund and Skovgaard (1973) found that the primary control on braiding is a width to depth (w/d) ratio greater than 50. Further stability analyses by Fredsoe (1978) and Fukuoka (1989) reconfirmed the requisite $w/d > 50$ criteria; Fredsoe found that the type of bedforms (flow resistance coefficients due to bars, ripples, dunes) were important and Fukuoka found that slope was critical.

Parker (1976) used stability analysis techniques to differentiate between meandering and braided regimes and derive a theoretical relationship for a system's stable number of braids. Parker's threshold for meandering verses braiding is summarized in textbox 2. Figure 4 displays a graphical representation of textbox 2 and the Copper River is plotted as a red star. Parker's results contradict the traditional hypothesis that braiding is caused by sediment loads in excess of

Textbox 2: Theoretical stability analysis of Parker (1976)

$$\varepsilon^* = \frac{s^* w}{\pi^* F^* d} \quad (3)$$

$$s = F^* d / w \quad (\text{simplified form}) \quad (4)$$

s = slope, w = channel width, d = flow depth

$F = v / \sqrt{g^* d}$ = Froude number, the ratio of flow inertial forces to gravitational forces

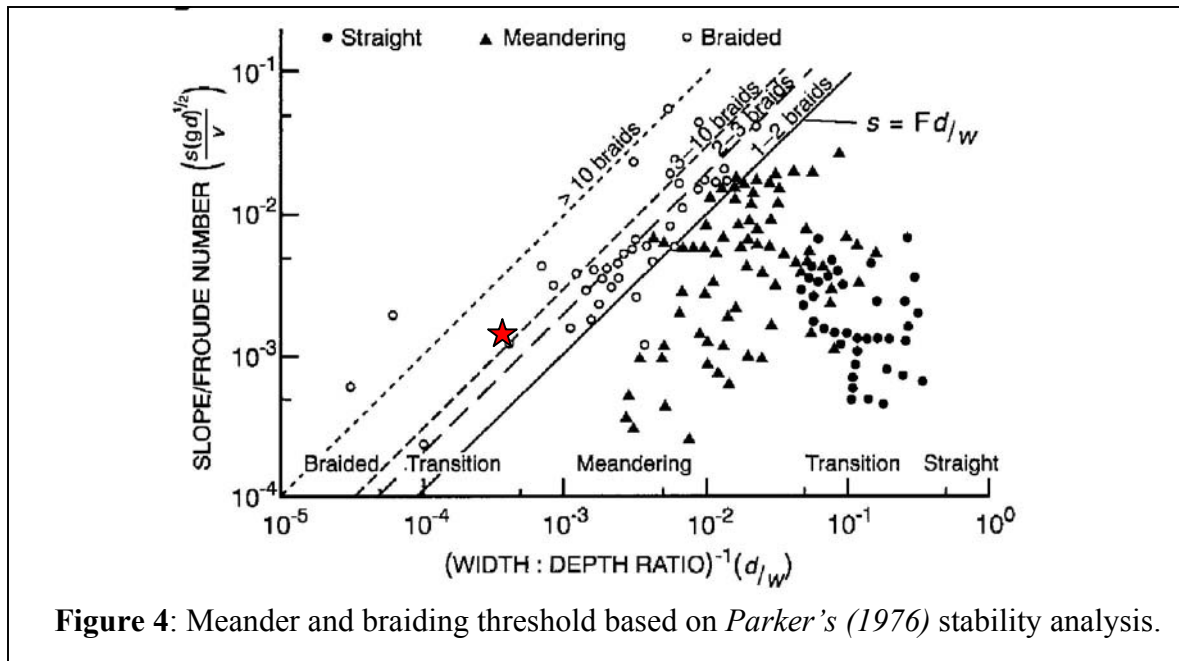
v = flow velocity, g = gravity, d = flow depth

$\varepsilon^* > 1$ braiding is favored, $\varepsilon^* < 1$ meandering is favored

ε^* = theoretical stable number of braids

transport capacity that result in bar deposition and general channel aggradation. The traditional view implies that braided rivers cannot be in equilibrium and that aggradation occurs until a higher equilibrium sloped is attained at which point braiding will cease. However, Parker's analysis indicates that rivers have a tendency to form bars and braids even when they are in

equilibrium. Flume experiments have also demonstrated that braiding can form in equilibrium conditions (Ashmore 1991). Bar formation will lead to braiding if the slope and w/d ratio are sufficiently high at formative discharges. Excessive sediment loads and subsequent aggradation can increase the tendency for braiding by increasing the slope and forcing the channel out its banks.



Sedimentary Mechanics

Both erosional and depositional processes can induce braiding. Based on flume studies, Ashmore (1991) found two depositional mechanisms that can induce braiding: central bar deposition and transverse bar conversion. Ashmore (1991) identifies two erosional mechanisms responsible for braiding: chute cutoff of solitary or alternating point bars and dissection of multiple bars. The mechanisms are primarily a response to the instability of rapidly changing flow and sediment transport during flow expansion (diverging flow zones). This instability gives rise to local aggradation of sediment (often by stalling of bedload sheets or migrating bars) in the flow expansion. There is typically a decline in flow competence (ability to transport sediment) during the development of the flow expansion, which leads to an accumulation of coarse particles at the onset of braiding. Once braiding is established all of the processes may occur simultaneously to promote further braiding. Central bar deposition is often referred to as the

primary or only mechanism of braiding; however, Ashmore (1991) found chute cutoff to be the primary mechanism and postulated that conditions of sediment mobility and channel morphology will determine which is prevalent.

Depositional Mechanisms: Central bar and transverse bar conversion (top figure 5)

The central bar mechanism initiates in an undivided channel with the deposition of a submerged central bar or a stalled bedload sheet (downstream migrating accumulation of mixed sediment with a coarse leading edge). Although the new bed form is only a few grain diameters thick, it quickly accretes portions of passing bedload sheets and grows laterally and in the headward direction. Local depth is reduced along the bar, but velocities tend to remain undiminished or increase, and some particles are transported along the length of the bar and deposited at the downstream end. Eventually the bar becomes large enough that the channel width is insufficient to remain stable, and the channel widens by cutting laterally against the banks and bar. As the channel widens, flow depth decreases and the bar begins to emerge above the water surface. Additional scour may occur along the channel flanks and the bar can emerge as an island. Islands are often stable enough to support vegetation, which leads to further sediment deposition. All of these processes can occur at a constant discharge.

Braided systems are likely to have multiple channels that diverge around bedforms or islands and reconnect flow paths at confluences. Confluence zones concentrate flow and erosive power, which often forms a pool due to local scour. As flow exits the pool it begins to expand and lose its transport capacity. Transverse unit bars are deposited downstream of the pool (figure 5). Flow is divided around the transverse bars, and additional sediment is deposited in the center of the bar as passing bedload sheets stall. Flow is divided around the emerging bar and scours the outer margins. The transverse bars continue to grow in a process analogous to the final stages of the central bar mechanism.

Erosional Mechanisms: Chute Cutoff and Multiple Bar Braiding (bottom figure 5)

Chute cutoffs occur by the development of channels across previously established alternate or point bars. Ashmore (1991) found that cutoffs typically occurred when bedload sheets accumulated on alternate bars and caused more flow to be directed across the bar. The redirected flow eventually scours a channel into the alternate bar (usually on the inner bank portion), and the remaining bar emerges as a medial braid bar. Multiple bar braiding results in a

morphology similar to that of the chute cutoff, but is developed from a different process. In multiple bar braiding, the formation of alternate bars is inhibited when flow depth is low and channel width is high. At the onset of multiple bar braiding, flow is concentrated into isolated, discontinuous chutes over submerged bars with bedload transport along the chutes. Eventually some of these chutes become well established, and local scour begins to occur. Bars are deposited downstream of the chutes, and flow is divided around the deposits. The pattern of chute channels and downstream deposition eventually develops into a channel network with bar complexes. Multiple bar braiding appears to be a special case that applies only to channels with very high width to depth ratios and is more common in sand-bedded versus gravel-bedded streams.

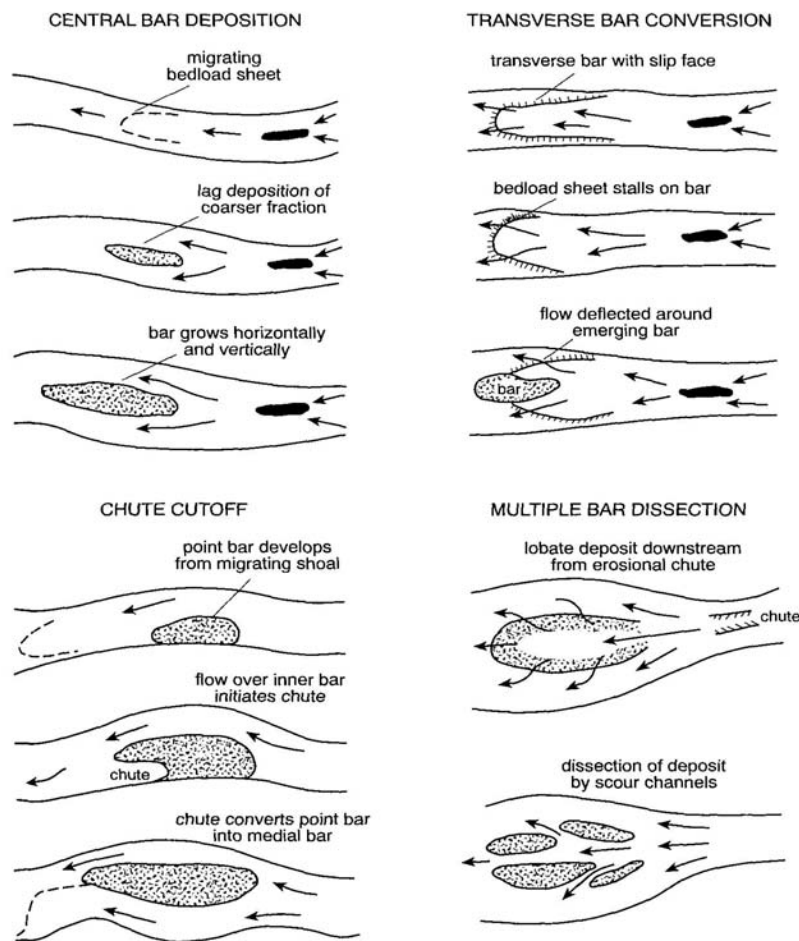


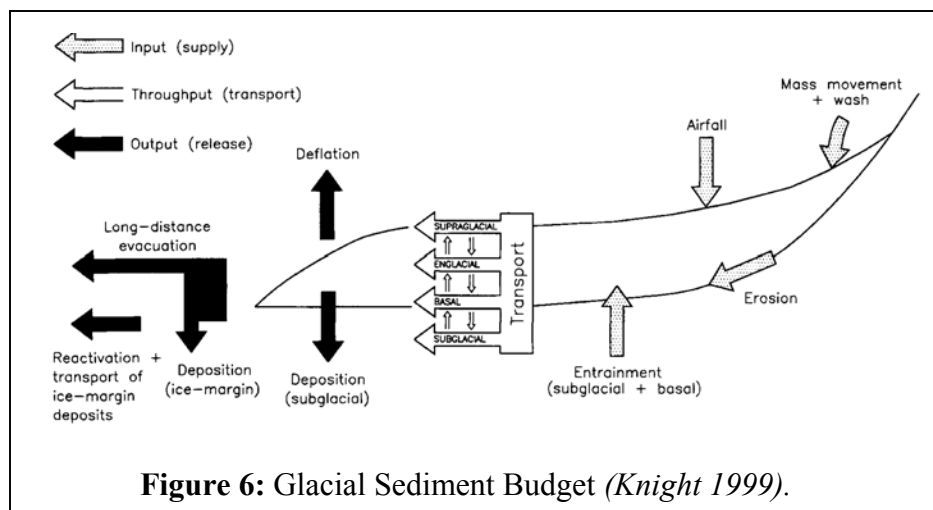
Figure 5.23 Mechanisms of braid development, based on flume experiments of Leopold and Wolman (1957) and Ashmore (1991a), and cartoons from Ferguson (1993).

Figure 5: Illustration of sedimentary mechanisms for braid bar initiation (Kighton 1998).

GLACIERS AND THEIR INFLUENCE ON SEDIMENT SUPPLY

Sediment Production and Discharge

Glacial systems typically have very high sediment fluxes. On an annual basis, glaciers advance downslope during the accumulation season and retreat during the ablation season. The continued movement of ice across the landscape causes substantial erosion by abrasion. Denudation rates beneath glaciers can vary from 0.01 mm/yr to more than 100 mm/yr (Knight 1999). Glaciers receive additional sediment inputs from hillslope mass wasting, sheet wash (erosion of hillslopes by rain), tributary inputs, and by entraining older sediment deposits. Figure 6 schematically illustrates glacial sediment inputs, transport paths, and outputs (collectively referred to as the sediment budget). Sediment budgets for glaciers are difficult to quantify because the majority of the sediment transfer system is inaccessible for detailed measurements. The locations and quantities of sediment input, storage, and output vary widely between glaciers. Warburton (1990) attempted to construct a sediment budget for the Bas Glacier d'Arolla in Switzerland. Warburton found that the glacier contributed about 23% of the total sediment load at the basin outlet and that 95% came from erosion of the valley floor.



Sediment production from glaciers varies widely depending on the processes that operate beneath them and the nature of the material over which they flow. Glaciers moving over unconsolidated sedimentary surfaces erode much faster than glaciers abrading bedrock. Quickly moving, warm glaciers tend to produce more sediment than slow, cold glaciers. Larger glaciers tend to vary little in their yearly sediment production, but small glaciers tend to have large

fluctuations in sediment production because periodic evacuation of high proportions of stored sediment can lead to exhaustion. Knight (1999, table 9.1) reported that the warm rapidly moving Johns Hopkins and Margerie glaciers in SE, Alaska have erosion rates of 47-60 mm/yr, an order of magnitude higher than selected glaciers in Europe and Asia. Glaciers in the Copper River Basin could be expected to have similar high erosion rates.

Hallet et al. (1996) found that basins with extensive glaciation produced much higher loads of sediment. Sediment discharge that takes place in a proglacial region (area beyond the terminal glacial edge) depends primarily on the amount of sediment made available by the glacier and from marginal zones such as moraines and screes, and the amount of meltwater available to transport sediment. Sediment discharge from a glacier is not necessarily reflected in downstream proglacial fluvial sediment fluxes. The proglacial fluvial sediment budget varies with sediment production and with periodic storage and release of sediment by traps at the glacier margin. The major factors controlling sediment throughput past the marginal zone and into watercourses are local topography and meltwater routing. Glaciers that terminate beyond topographic obstructions, such as end moraines, can release sediment from the ice margin directly to the proglacial zone. Glaciers ending behind fringing moraines are likely to have their sediment stored in traps such as lakes and other closed basins. Whether the topography underlying the glacier concentrates meltwater into channels or diffuses water will have a major impact on transport capacity. Moraine ridges are common glacial mechanisms for focusing meltwater discharge and increasing fluvial transport capacity. Sediment discharge from a glacier to a stream is a complex and locally variable balance between fluvial transport capacity, sediment supply, and the topographic pathways between the glacial and proglacial environment.

Glacier Flux and Downstream Impact

A glacier's state of flux (advancing, stationary, or retreating) is an important variable in controlling sediment and meltwater supply. Glacial state of flux is defined by whether the glacier has a net gain (advance) or loss (retreat) in ice volume per year. Sediment discharge varies with glacier advance and retreat due to the interactions of glaciers with fringing sediment sources and to variations in meltwater discharge and flow competence. Sediment from advancing glaciers is less likely to encounter topographic obstructions or sediment traps and should have a higher likelihood of reaching the channel. However, if previous glacial episodes

advanced beyond the current stage, then moraines and other sediment traps are likely in the proglacial zone. Sediment from retreating glaciers is more likely to be influenced by recently exposed moraines and deposit prior to reaching the channel. Advancing glaciers can entrain sediment from proglacial moraines and the sediment stored behind them. However, meltwater from retreating glaciers has access to recently exposed moraines.

The role of glacial flux on channel aggradation or degradation is unclear and both may be associated with glacial retreat or advance. Some authors claim that maximum channel aggradation occurs during glacial advance due to enhanced glacial erosion (as advancing ice abrades the valley floor) supplying high sediment loads to be deposited by meltwater streams (Schumm 1977). Other authors suggest aggradation occurs during glacial retreat when greater quantities of debris are made available to meltwater streams by the exposure of deposits as ice retreats and meltwater discharge increases from the mass melting of ice (Flint 1971). Maizels (1979) studied the Bossons Glacier in France over a six-year period as the glacier was advancing. The longitudinal profile of the Bosson valley train experienced degradation during the first two years of Maizels's study, but the channel showed extensive aggradation for the following four

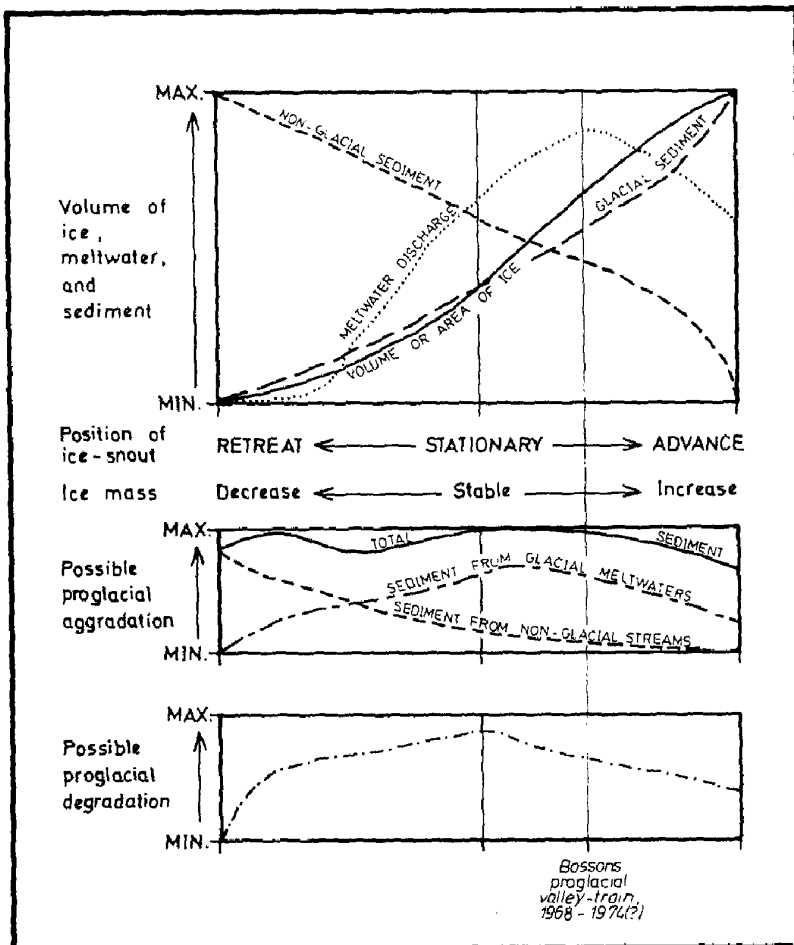


Figure 7: Diagram of potential aggradation or degradation versus glacial retreat and advance (Maizels 1979).

years as the advancing glacier accessed large quantities of proglacial outwash and moraine debris. Maizels developed a tentative model (Figure 7) illustrating how aggradation or degradation could result from glacial advance or retreat. The state of glacial flux on sediment inputs to the proglacial fluvial zone varies with local conditions and needs to be analyzed on an individual case basis.

Glacier Dammed Lakes and Outburst Floods

Glaciers often flow across the mouths of adjacent valleys and effectively dam the tributaries and form large lakes. Glacial ice dams are subject to repeated failure, usually as the lake level rises to a point sufficient to float the ice dam. Dam failures result in major outburst floods known as jökulhlaups [refer to (De Paoli, 2002) in this volume for more detail]. Jökulhlaups and ice dams are common in heavily glaciated regions and they have significant impacts on glacial and proglacial stream channel sediment fluxes. Due to their elevated water stage, jökulhlaups are able to deliver large sediment volumes as they access sediment deposits and glacial moraines located significant distances from the channel. Jökulhlaups also have extremely high sediment transport capacity due to flood level discharges. If they recur frequently, jökulhlaups can become the formative discharge for a system and dictate a channel's geometry, textural composition and degree of braiding.

COPPER RIVER ANALYSIS

The study reach along the Copper River ranges from the Copper River Delta to the confluence with the Chitina (figure 8). From the Chitina confluence the study reach extends east to the Kennicott Glacier along the Chitina, Nizina, and Kennicott Rivers. The headwaters of the Copper River basin are bounded by three mountain ranges: the Alaska Range to the north, the Wrangell-St. Elias Mountains to the east, and the Talkeetna Mountains to the west. The Chugach Mountains divide the Copper River basin into two distinct climate types. The upper part of the basin lies to the north of the Chugach and has a cold and arid climate typical of interior Alaska. The Upper Copper River drains the western portion of the basin, and the Chitina River is the primary tributary to the east. At the confluence with the Chitina, the Copper River makes a distinct bend to the south where it begins to bisect the Chugach Range. South of the

Chugach Range, the Copper River is a maritime climate with moderate temperatures and high rainfall. Glaciers and glacial lakes have at one time or another covered the majority of the Copper River watershed. Glaciation is still the primary geomorphic force; in 1995 approximately 18 percent of the Copper River watershed consisted of glaciers (Brabets 1997). The two major zones of present day glaciation in the Copper River basin are in the headwater regions and the southern front of the Chugach Range near coastal proximities (Figure 8). Glaciers in the Copper basin are currently in a state of retreat.

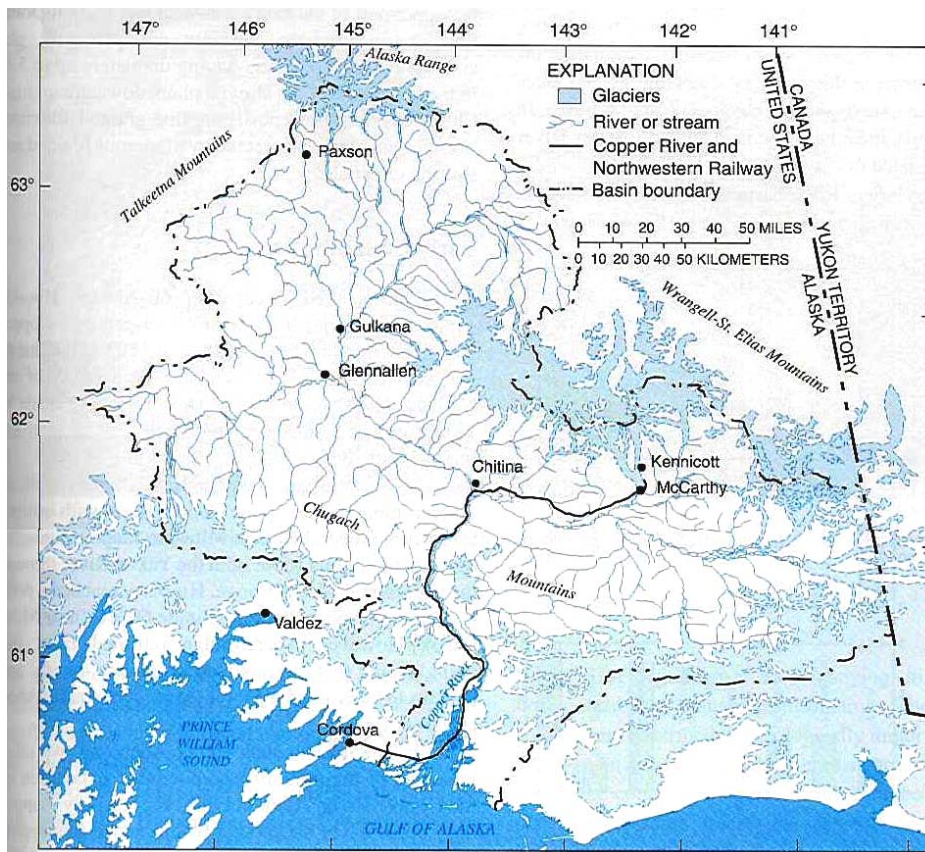


Figure 8: Copper River Basin (outlined in hatch pattern), shaded light blue represents glacial areas (Brabets 1997).

Geomorphic Properties

Discharge

The mean annual discharge of the Copper River is approximately 1,625 m³/s. Although the Copper River watershed is the sixth largest in Alaska, it has the second largest mean annual discharge and average discharge per square mile. The formative discharge on the Copper River is approximately 5,950 m³/s, if a flow event with a 1.5-2 year recurrence interval is assumed representative of the formative discharge [refer to (Bowersox, 2002) in this volume for more detail]. The maximum recorded discharge on the Copper River is 13,300 m³/s for a 45-year

period of record. Jökulhlaups are known to have flooded the Lower Copper River, below Miles Lake. Historical jökulhlaups are inferred along the remaining channel lengths in the study reach from Miles Lake up to the Kenicott Glacier (Post and Mayo 1971). Miles Glacier currently forms a glacier dam by impounding meltwater from Van Cleve Glacier. Van Cleve has a break out flood about every seven years, although, the increase in discharge is relatively small, estimated to be a 1.5-2 year flood event [refer to (De Paoli, 2002) in this volume for more detail].

Sediment Load

The predominance of sediment transport in the Copper River initiates in the spring with the snow melt and is maintained throughout the summer by glacial meltwater and rain. The suspended sediment load (comprised of silt and clay) of the Copper River averaged 69 million tons per year, ranging from 63-73 million tons, over a three-year period (Brabets 1997). The Copper River has a total suspended sediment load similar to the Yukon (the largest basin in Alaska); however, the Copper River has a suspended yield (tons per square mile) that is a factor of 2 higher than other Alaska rivers. The USGS measured bedload discharge at four bridges below Miles Lake and found that bedload approximated about 5% of the total sediment load. Bedload measurements between the 75th and 25th percentiles ranged from 1,000-8,000 tons per day. The median grain size diameter (D₅₀) of the bedload samples ranged from 0.5-16.0 mm. A summary of discharge, sediment transport, and other geomorphic properties is listed in *Table 1*.

Table 1: *Geomorphic Properties of the Copper River (all values are for the lower Copper River except where noted.)*

Watershed Area ^a	26,500 mi ²
Mean Annual Discharge ^a	1,625 m ³ /s
Approximate 2 yr flood event ^a	5,950 m ³ /s
Max Discharge ^a	13,300 m ³ /s
Slope	
	Copper River 0.0009
	Chitina River 0.0018
	Nizina River 0.0040
	Kennicott River 0.0077
Froude Number	0.5-0.8
Width to Depth Ratio	250-500
Mean Suspended Sediment Load ^a	69 million tons per year
Bedload Transport Rates ^a	1,000-8,000 tons per day
D ₅₀ of bedload samples ^a	0.8-16 mm
Mean D ₅₀ of bed material samples ^a	40 mm

^a= data from *Brabets* (1997) report on the Lower Copper River

The current sediment budget of the Copper River is adjusting to the external forces of tectonic uplift and glacial retreat. The Chugach-St Elias region of southern Alaska is an actively deforming glaciated orogenic (mountain) belt. Estimated uplift rates for the region are ~4 cm/yr from tectonic uplift and glacial rebound (glacial melt removes sufficient ice mass that allows the Earth's crust to flex outward) associated with glacial retreat over the last 100+ years (Jaeger 2001). Continued uplift of the region will have significant effects on the Copper River's sediment supply and longitudinal profile. The Copper River and much of the southern Alaskan coast experienced massive uplift from the magnitude 9.2 earthquake on March 27, 1964. The Copper River Delta was uplifted 2 m from the quake (Brabets 1997). The 2 m uplift is a dramatic change in longitudinal profile for a system that has a 0.001 slope in its lower reaches. The uplifted profile will impose a local reduction in channel slope and transport capacity, which in the short term should lead to channel aggradation. Aggradation should persist until local slopes steepen sufficiently to allow the Copper to incise back to its graded profile. The retreating flux state of glaciers in the basin will also impact sediment discharge in the basin. The Miles glacier lost an area of about 9.3 mi² between 1910 and 1950, and from 1950 to 1991 the glacier lost an area of 2.0 mi² (Brabets 1997). The total impact on channel aggradation is unclear, but the retreating glaciers should increase discharge due to ice melt, and new sources of sediment will be exposed as glaciers retreat.

Slope and Braiding Parameters

The longitudinal profile of the study along the Copper River and its tributaries can be seen in Figure 9. Channel slopes are sufficiently steep for their respective discharges to induce braiding because the Copper River is heavily braided along the entire study reach. According to data on USGS 7.5' minute quadrangles the Kennicott River begins braiding directly at the toe of the Kennicott glacier. Using data from *Table 1* the Copper River can be plotted on various braiding/threshold regime diagrams. Using equation (3) the Copper River's $\epsilon^* = 2.3$ (clear tendency towards braiding) and is plotted in Figure 4 as a red star. However, using ϵ^* as a proxy for number of braids is a clear underestimate for the Copper River based on USGS quad maps. For Ferguson's (1984) regime diagram, equation (2), the Copper River plots as expected in the braided regime for its median grain size (see Figure 3 red star).

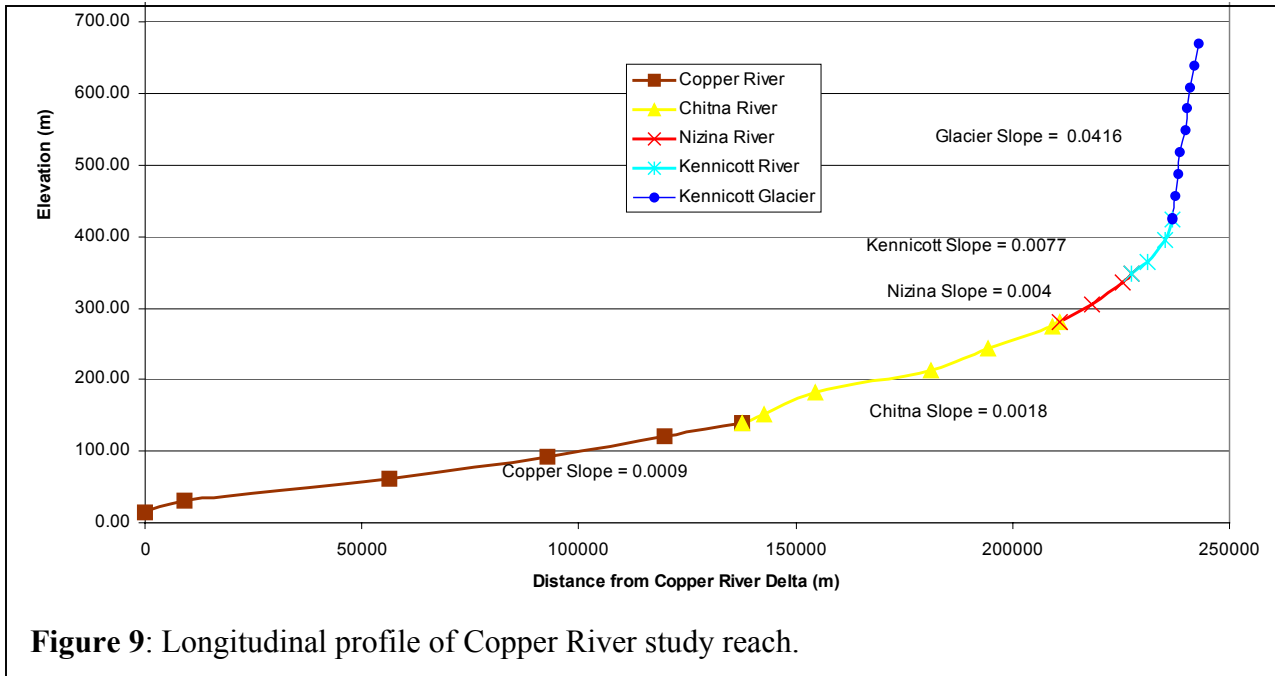


Figure 9: Longitudinal profile of Copper River study reach.

SUMMARY

The Copper River Basin is a glacially dominated system with some of the highest discharge and suspended sediment loads per basin area in Alaska. The Copper River is currently adjusting to rapid uplift from the 1964 earthquake and a retreating glacial flux. The Copper River is a prime example of a glacial, braided system and affords an excellent opportunity for further investigation on braiding in a glacial environment. Glacial systems characteristically have high sediment loads, discharge, and slopes, which usually produce a braided river. Sediment discharge from a glacier to a proglacial river system is a complex and locally variable balance between fluvial transport capacity, sediment supply, topographic pathways between the glacial and proglacial environment, and glacial state of flux. Although why a river forms a braided channel is not completely understood, a high bedload supply, a width to depth ratio greater than 50, erodible bank material, and high stream power are essential factors for a braided system.

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